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GEOMATRIX

FINAL REPORT

PART 1

SEISMIC SOURCE MODELS, RECURRENCE MODELS, AND GROUND MOTION ATTENUATION MODELS SEISMIC HAZARD IN SOUTHERN ONTARIO

Prepared for:

Atomic Energy Control Board
280 Slater Street, 7th Floor
Ottawa, Ontario K1P 5S9, Canada

AECB Contract No. 87055-5-5072/001/SS

Prepared by:

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FINAL REPORT
PART 1
SEISMIC SOURCE MODELS, RECURRENCE MODELS AND GROUND MOTION ATTENUATION MODELS SEISMIC HAZARD IN SOUTHERN ONTARIO

1.0 INTRODUCTION

This report is Part 1 of the three-part final report on the Seismic Hazard of Southern Ontario study conducted for the Atomic Energy Control Board of Canada. Part 1 entails the characterization of all seismic sources that could affect the ground motions at the Darlington and Pickering sites, and the selection of ground motion attenuation relationships for use in the probabilistic seismic hazard analysis. Part 2 will include the probabilistic seismic hazard analysis and sensitivity analysis to identify dominant contributors to the hazard results. Part 3 will include recommendations regarding the most significant technical issues to the seismic hazard and the data collection and other activities that could address these issues.

The purpose of this report is to present the interpretations of potential seismic sources and ground motion attenuation, developed on the basis of available literature and other information. A draft of this report was submitted to the Multi-disciplinary Panel for review and, after revision, to a number of external peer reviewers. In this final report we have addressed and incorporated to the extent possible the comments of the peer reviewers. The goal of this report is to ensure that:

- A full range of hypotheses have been included
- The tectonic hypotheses have been fairly and properly represented
- The seismogenic potential of each potentially active source has been properly quantified in light of available data

- The logic structure for evaluation of alternative source interpretations is reasonable
- Regional source zones are appropriate and their basis clear
- Approaches to assessing maximum magnitudes and earthquake recurrence are appropriate
- Ground motion attenuation and associated uncertainties have been properly represented.

Before discussing the methods and activities conducted for this study, we first provide an introductory discussion of probabilistic seismic hazard analysis (PSHA). The purpose is to define general concepts and terms, explain the ways that uncertainties are incorporated into the analysis, and to lay out the logic structure used for the seismic source characterization.

1.1 SEISMIC HAZARD ASSESSMENT METHODOLOGY

The evaluation of seismic hazards at the Darlington and Pickering sites is to be based on a probabilistic seismic hazard analysis (PSHA). The methodology for PSHA was first proposed by Cornell (1968, 1971). Since that time there has been significant progress in scientific understanding of the earthquake process and in the techniques for evaluation of the relevant seismological, geological, and geophysical data. This has led to improved methods of PSHA that provide greater flexibility in representing the distribution of future earthquakes in time, space, and size, and the resulting levels of ground motion that may be produced at a site; and improved methods for estimating the parameters of these distributions. PSHA studies now form the basis for the seismic provisions of building codes in Canada and the United States and are used to assess seismic risk at critical facilities, such as nuclear power plants, in both countries.

The basic components of the PSHA formulation remain the same as originally proposed by Cornell (1968, 1971) and are shown schematically on Figure 1-1. The components are as follows:

(1) The sources of potentially damaging future earthquakes are identified. A *seismic source* in a PSHA is a region having relatively uniform seismicity characteristics. Seismic sources are typically identified based on geologic and observed seismicity data. The types of sources typically identified are specific geologic structures, such as faults, and areal source zones representing tectonic provinces or zones of seismicity.

(2) The frequency of earthquake occurrence—or earthquake recurrence—in each source is assessed and expressed as a frequency vs. magnitude relationship. This includes an evaluation of the maximum event that a source can produce (M1 and M2 in Figure 1-1). The probability distribution of distance from individual earthquakes to the site is also defined.

(3) Appropriate ground motion models are selected for estimating site ground motions from each of the sources. The estimates include both the expected levels of ground motion and the variation about the expected value that any single recording may exhibit.

(4) Relationships between ground motion level and probability (frequency) of exceedance are developed. These define what are called *hazard curves*. Hazard curves for various spectral frequencies are obtained by repeating the PSHA calculation using attenuation relationships for each spectral frequency. Specific design ground motion levels for site evaluation can then be chosen by selecting an appropriate probability of exceedance. The corresponding ground motion levels at each spectral frequency form what is called an *equal-hazard* response spectrum.

The PSHA methodology outlined on Figure 1-1 is formulated to represent the randomness inherent in the natural phenomena of earthquake generation and seismic wave propagation. The randomness in the physical process has come to be called *aleatory* uncertainty (SSHAC, 1996). In all assessments of the effects of rare phenomena one is faced with uncertainty in selecting the appropriate models and model parameters arising from limited data and/or alternative interpretations of the available data. This uncertainty in knowledge has come to be called *epistemic* uncertainty. The focus of this study is on developing a quantitative description of the epistemic uncertainty.

The uncertainty assessment in this study is performed using a logic tree methodology. The logic tree formulation for seismic hazard analysis (Kulkarni and others, 1984; Coppersmith and Youngs, 1986; Electric Power Research Institute, 1986; National Research Council, 1988) involves setting out the sequence of assessments that must be made in order to perform the analysis and then addressing the uncertainties in each of these assessments in a sequential manner. Thus, it provides a convenient approach for breaking a large, complex assessment into a sequence of smaller, simpler components that can be more easily addressed.

The general structure of the logic tree used in this study is shown on Figure 1-2. The logic tree is composed of a series of nodes and branches. Each node represents an assessment of a state of nature or an input parameter value that must be made to perform the analysis. Each branch leading from the node represents one possible discrete alternative for the state of nature or parameter value being addressed. If the variable in question is continuous, it can be discretized at a suitable increment. The branches at each node are intended to represent mutually exclusive and collectively exhaustive states of the input parameter. In practice, a sufficient number of branches are placed at a given node to adequately represent the uncertainty in the parameter estimation.

Probabilities are assigned to each branch that represent the relative likelihood or degree of belief that the branch represents the correct value or state of the input parameter. These probabilities are assessed conditional on the assumption that all the branches leading to that node represent the true state of the preceding parameters. Because they are conditional probabilities for an assumed mutually exclusive and collectively exhaustive set of values, the sum of the conditional probabilities at each node is unity. The probabilities are usually based on scientific judgments because the available data are too limited to allow for objective statistical analysis, and because scientific judgment is needed to weigh alternative interpretations of the available data. The logic tree approach simplifies these assessments because the uncertainty in a single parameter is considered individually with all other assessments leading up to that parameter assessment assumed to be known with certainty. Thus, the nodes of the logic tree are sequenced to provide for the conditional aspects or

dependencies among the parameters and to provide a logical progression of assumptions from the general to the specific in defining the input parameters for an evaluation.

In most cases, the probabilities (relative weights) assigned to the branches at a node are in units of tenths, unless there is a basis for more fine-scale resolution. The weights represent one of two types of probability assessments. In the first, the branches at a node of the logic tree define the range of possible parameter values and the associated weights assigned to the branches define the probability distribution. For example, estimates of the slip rate on a fault is uncertain because of uncertainties in the amount of displacement of a particular geologic unit across the fault and the age of the unit. The resulting slip rate is usually represented by a preferred value and a range of higher and lower values, similar to a normal or log normal statistical distribution. This type of distribution can be represented by three (or more) branches of a logic tree. Keefer and Bodily (1983) have shown that most distributions can be reliably represented by three values: the median estimate (50^{th} percentile) with a weight of 0.63 and a higher and lower value, each with weights of 0.185, that represent the 5^{th} and 95^{th} percentiles (± 1.65 standard deviations for a normal distribution). Another four-point representation is described in EPRI (1993).

In some instances, this type of uncertainty in parameter assessment can be estimated using formal statistical techniques. In these cases, continuous parameter distributions developed from statistical estimation procedures can be discretized for use in a logic tree formulation. Such an approach will be used to develop discrete distributions for earthquake recurrence parameters and maximum magnitudes in Section 5.0.

A second type of probability assessment to which logic trees are particularly well suited is in indicating a relative preference for, or degree of belief in, two (or more) alternative hypotheses. For example, the sense of slip on a fault may be uncertain, two possible alternatives might be strike-slip or reverse-slip. Based on the pertinent data, a relative preference for these alternatives can be expressed by the logic tree weights. A very strong preference for one over the other is usually represented by weights such as 0.9 and 0.1 for the two alternatives. If there is no preference for either hypothesis, they are assigned equal

weights (0.5 and 0.5 for two hypotheses). Increasing the weight assigned to one of the alternatives from 0.5 to 0.9 (or more) reflects an increasing preference for that alternative. Because the relative weights are ultimately scientific judgments based on available information, it is important to document the data and interpretations that led to the assessment of parameter values and led to assignment of relative weights in order that the process can be reviewed by others.

Figure 1-2 lays out the logic tree structure that will be used in this PSHA. The logic tree is schematic in that only the general levels of the logic tree are indicated. Sections 4 , 5, and 6 provide more detailed logic trees.

The first level of the logic tree addresses the uncertainty in defining the appropriate ground motion attenuation model. Although uncertainty in identifying and characterizing seismic sources is the focus of this study, the uncertainty in ground motion characterization will be included in the analysis in order to indicate its relative importance in the overall uncertainty in the seismic hazard. The attenuation models that will be used in the PSHA are discussed in Section 6.

The second level of the generalized logic tree addresses the uncertainty in the appropriate approach for regional seismic source zonation. The importance of this level of the logic tree is that alternative approaches to regional seismic zonation may lead to the definition of different seismic sources. Section 3 discusses the various ways that have been used to identify seismic sources in the region and describes the basis for the seismic sources. Section 4 lays out the complete logic tree structure used to model the uncertainty in seismic source identification. As described in Section 4, there are several levels of assessments used to identify regional seismic sources and the single level labeled *Regional Source Zonation Approach* on Figure 1-2 is expanded into multiple levels in the complete logic tree.

At this point the logic tree is expanded into subtrees, one for each seismic source identified following a particular zonation approach. [The vertical line without a node (dot) on Figure 1-2 denotes summation of the hazard from multiple independent sources.] As discussed in

Section 3, two types of seismic sources are defined in this study, regional areal sources and local sources. The regional sources are identified by 1A and 1B to indicate that different sources may be defined using different zonation approaches. The local sources are given the same designation to indicate that their definition is independent of the regional zonation approach.

The next level of uncertainty addressed is the likelihood that the individual local sources are "seismogenic" (seismogenic is defined in Section 3.2.1). If a local source is seismogenic, then it is considered a discrete seismic source in addition to the regional source zone. If not, then only the regional source zone is present. This assessment is only applied to the local sources, the regional source zones are considered to be seismogenic with certainty (probability 1.0).

The final four levels of the logic tree address the assessment of maximum magnitude and seismicity rate parameters. For each there is an assessment of the appropriate approach and then an assessment of the parameters following the selected approach. The approaches that will be used for assessment of maximum magnitude and seismicity rate parameters are discussed in Section 5.

1.2 METHODS USED IN THIS ANALYSIS

The focus of the seismic source characterization has been to identify and incorporate a full range of hypotheses and models related to the seismic environment of the Darlington and Pickering site region. In addition to compiling the alternative hypotheses, we have also attempted to translate those hypotheses into a format that is suitable for input into a probabilistic seismic hazard analysis. For example, one tectonic hypothesis that is considered is the possibility that a branch of the St. Lawrence Rift extends southwest of the Ottawa Graben into the Lake Ontario and Lake Erie region. To incorporate this hypothesis, a seismic source zone is identified whose boundaries represent the postulated structural boundaries of the zone—and their associated uncertainties—as depicted in the available literature. Further, the credibility of this hypothesis relative to other hypotheses (e.g., no southwest extension but only a regional source zone representing the Grenville province, or a regional source zone including areas of elevated observed seismicity) is evaluated and quantified. By establishing

the relationships among the alternative seismic source interpretations and their relative weights, it is possible to combine all of the source interpretations into a seismic hazard analysis.

The basic steps followed in the seismic source interpretations were the following:

(1) review of available literature regarding seismotectonic hypotheses

(2) definition of the geometries of alternative seismic sources

(3) characterization of the seismogenic potential (probability of activity) of each seismic source

(4) identification of potential methods for assessing maximum magnitudes and earthquake recurrence for each source.

The results of each of these steps is discussed in detail in subsequent parts of this report. Below we summarize the activities associated with each step.

1.2.1 Literature Review

The literature review included the recent pertinent geological, geophysical, and seismologic professional literature; the summary of the AECB Workshop on Seismic Hazard in Southern Ontario; AECB contractor reports; reports of studies conducted for Ontario Hydro; and unpublished reports of recent or ongoing studies (e.g., USNRC study of methods for evaluating fault earthquake potential; EPRI study of eastern North America ground motion models). The review included the AECB "Data Base", which provides one-page summaries of all pertinent AECB reports and publications regarding the earthquake environment of southern Ontario. In some cases, individual researchers were contacted to provide additional clarity to their interpretations or to update their published interpretations. For example, Dr. John Adams was consulted about the basis for the "geological" and "seismological" zones provided in his 1995 paper. The information gathered in the literature review task provides a basis for the section of this report on the Regional Tectonic and Geologic Setting, as well as the technical basis for the regional and local seismic sources.

1.2.2 Seismic Source Locations and Geometries

Seismic source interpretations consist of *regional* seismic source zones and *local* seismic sources. For this study, regional seismic sources are interpreted within about 300 km of the sites; local seismic sources are interpreted within about 100 km of the sites. The geometries of the regional seismic sources come primarily from the locations of regional-scale tectonic boundaries, significant zones of observed seismicity, and combinations of the two. In most cases the regional "geological" zones proposed by the Geological Survey of Canada (Adams and others, 1995) provide the basis for the regional zones, although there are some differences. A concerted attempt has been made to capture the impact of alternative hypotheses on regional source configurations. Examples are differences in the location of the boundary between the Grenville Province zone and Iapetan rifted crust.

Local seismic sources are defined based on existing tectonic hypotheses, concentrations of observed seismicity, and geologic data. Because the probabilistic seismic hazard analysis that will be calculated for this project will be *site-specific* rather than regional (as developed by the GSC for all of Canada), we have devoted considerable effort toward properly representing a variety of tectonic hypotheses within the local site region. For example, potential seismic sources include pronounced geophysical lineaments (e.g., the Niagara-Pickering linear zone), interpreted regional fault zones (e.g., the Hamilton-Presqu'ile fault zone), local zones of seismicity (e.g., the Toronto-Hamilton seismic zone), and local faults (e.g., the Rouge River fault zone). Rather than deem any particular hypothesis insignificant to hazard (before the calculations are made), we have attempted at this time to include all potentially significant hypotheses. The subsequent hazard calculations and sensitivity analyses will provide a means of examining the relative importance of alternative hypotheses.

In order to incorporate all of the local seismic sources in the hazard analysis, it is necessary to consider the dependencies among the hypotheses. For example, two interpretations might represent mutually exclusive hypotheses; that is, if one seismic source is assumed to be active (seismogenic, as defined below), then another source might not be active. Alternatively, the activity of two sources might be independent. The methodology followed for establishing

source dependencies follows that of the EPRI seismic hazard study for the eastern U.S. (EPRI, 1986; 1989).

1.2.3 Seismogenic Potential

Each potential local seismic source is evaluated according to whether or not it is active or "seismogenic," where seismogenic is defined as capable of generating moderate-to-large earthquakes ($m_b \geq 5$) in the present tectonic regime. (All regional seismic sources are assumed to be seismogenic with certainty). This assessment methodology is very similar to that carried out in the EPRI study (Stepp and others, 1985). It is designed to make explicit the several steps involved in evaluating whether or not a particular tectonic feature is seismogenic. First the criteria are defined for assessing the seismogenic potential of each tectonic feature. Second, based on these definitions, each criterion is evaluated as to how diagnostic it is in assessing seismogenic potential. Third, each tectonic feature is evaluated relative to the various criteria. For example, the feature-specific data sets are used to assess whether or not the tectonic feature is associated with seismicity. Fourth, the probability that each feature is seismogenic is calculated based on the feature-specific assessment and the relative weights assigned to each criterion. The advantage of the approach used in this study—as opposed to a direct assessment of the probability of activity—is that the assessment is more transparent and the use of the available data is documented.

1.2.4 Maximum Magnitudes and Earthquake Recurrence

The final step in the evaluation of seismic sources in this study is the identification of the methods and procedures that can be used to evaluate the maximum magnitude and earthquake recurrence rate for each seismic source. The goal is to capture properly the uncertainties inherent in these assessments, both those related to alternative *models* that might be appropriate for evaluating maximum magnitude and recurrence and those related to alternative *parameter values* that might be appropriate for each model. In this report we assess possible alternative models that we believe are appropriate for the evaluation. In the Part 2 report, we will incorporate a range of parameter values, conduct calculations, and arrive at actual maximum magnitude probability distributions and source-specific recurrence relationships that properly incorporate the full range of modeling and parameter uncertainties.

1.3 TRIGGERED SEISMICITY

Earthquakes triggered by mechanical changes in the subsurface induced by human activities are a potential consideration in seismic hazard analysis. Human activities known to have triggered earthquakes include: reservoir impounding; deep wells either injecting or extracting fluids (waste disposal; hydrocarbon extraction; brine extraction); deep mines; large quarries; and underground explosions. Although triggered earthquakes may represent a significant issue relative to the assessment of seismic hazard (e.g., it may affect the timing or recurrence of earthquakes relative to a "natural" or background rate), we explicitly are not considering the issue in this analysis. Triggered earthquakes represent a distinct component of the seismicity and the hazard from these earthquakes needs to be considered separately from the hazard from natural seismicity. The differences between natural and triggered earthquakes encompass general characteristics as well as the behavior of specific sources. The component of the seismicity that is triggered reflects the recent growth of industrial activities; it is probably negligible over most of the 19th century, and has increased since. Thus, the overall temporal distribution of triggered and natural seismicity are different. Furthermore, triggered earthquakes occur where natural conditions are sufficiently close to failure and where human intervention alter these conditions by a sufficient amount. Since human factors are not evenly distributed, the overall spatial distribution of triggered and natural earthquakes are generally different. We therefore acknowledge the potential impact that triggered may have on the seismic hazard assessment, but assume that its contribution to the hazard will be characterized separately.

1.4 CONTENT OF REPORT

This report includes the following: Section 2 presents the regional geologic and tectonic setting from the standpoint of important seismotectonic features, their geologic history, and regional seismicity. Section 3 includes a discussion of the seismic sources that are identified for this analysis. First, the regional seismic sources are discussed, followed by the local sources. A key part of the assessment of local sources is the evaluation of their seismogenic potential. Section 4 presents issues related to how the various seismic sources are modeled for purposes of probabilistic seismic hazard analysis (PSHA). In Section 5 the various approaches that are proposed for evaluating maximum earthquake magnitudes and earthquake recurrence

relationships are discussed. Section 6 summarizes the approach used to capture the uncertainty in ground motion attenuation models that are believed to be appropriate for use in the seismic hazard study. Finally, Section 7 presents the references cited in the report.

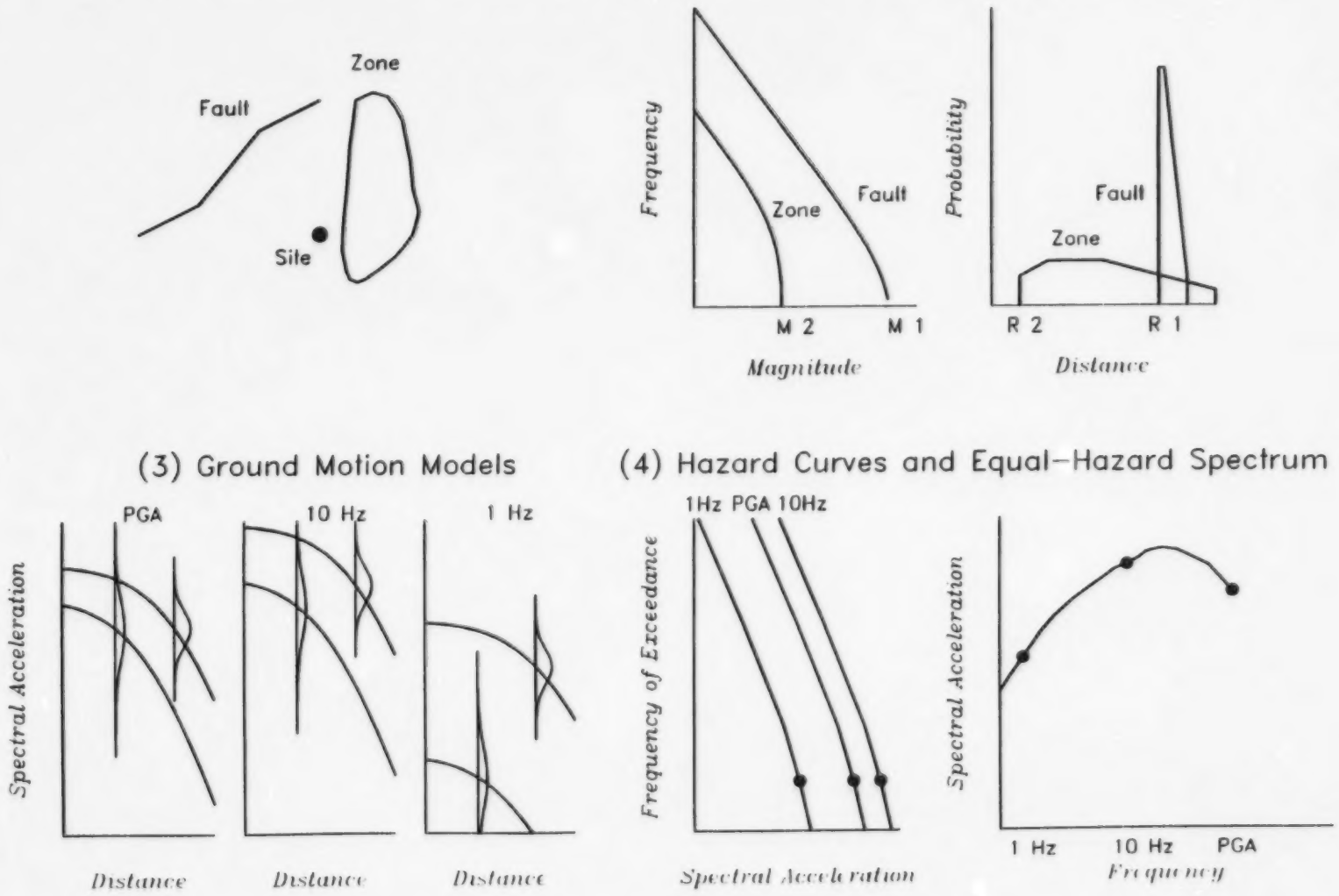


Figure 1-1 Schematic diagram of the components of PSHA

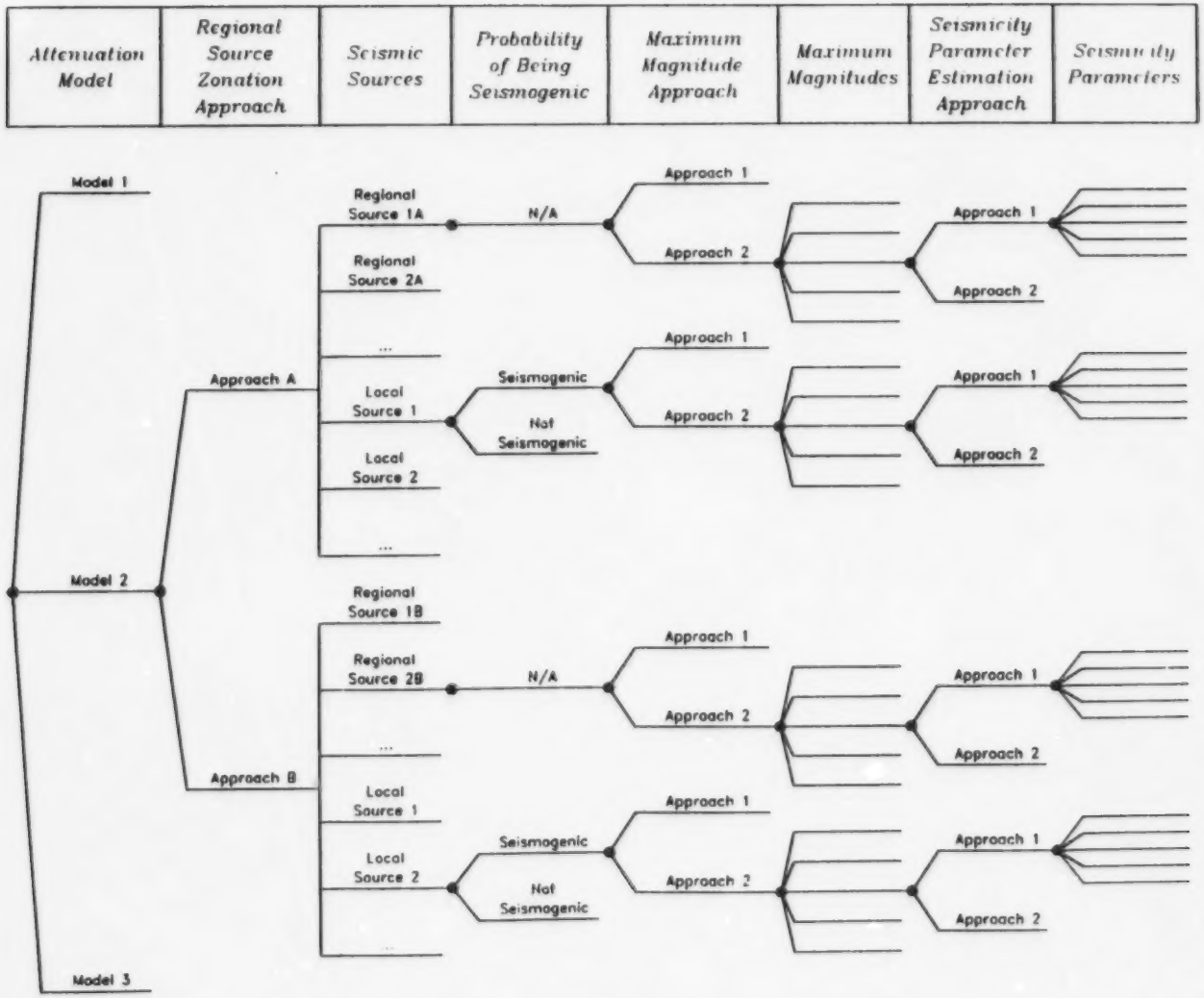


Figure 1-2 General logic tree structure to be used in PSHA

2.0 TECTONIC AND GEOLOGIC SETTING

2.1 REGIONAL TECTONIC SETTING

The Pickering and Darlington sites lie within the tectonically stable interior of the North American continent, which is characterized by low rates of historical seismicity (Wheeler and Johnston, 1992; Johnston and others, 1994). This stable continental region comprises non-extended middle Proterozoic Grenville basement (Bickford and others, 1986; Johnston and others, 1994) and overlying lower Paleozoic shallow-water sedimentary strata of the Central Craton (Figure 2-1). Rocks of the approximately 1100 Ma Grenville Province include remobilized basement rocks of Superior (~ 2500 Ma) and Hudsonian (~ 1800 Ma) age, and consist mainly of gneisses and thick sequences of high-grade metasediments extensively intruded by granitic and anorthositic bodies (Taylor and Toksoz, 1982). The Grenville crustal structure is characterized by north-northeast-striking, relatively shallow east-southeast-dipping ductile thrust faults that developed at mid- to lower crustal depths during the middle Proterozoic Grenville orogeny (Green and others, 1988; Hanmer, 1988; Milkereit and others, 1992; Forsyth and others, 1994a,b; Zelt and others, 1994). Prominent north-northeast-trending geophysical anomalies associated with exposed Grenville structures extend southward beyond the Canadian Shield and beneath the unconformable lower Paleozoic cover rocks. Regional geologic maps (e.g., Ontario Geological Survey, 1991) indicate that the overlying Paleozoic rocks are, with few exceptions, relatively flat-lying and laterally continuous, indicating that no large-scale, major faulting has occurred in the region since they were deposited.

The notable exception to the lack of regional-scale faulting in southeastern Canada occurs within the St. Lawrence rift system (Kumarapeli and Saull, 1966), which is a remnant of the late Proterozoic-early Paleozoic Iapetan passive margin (Wheeler, 1996a). The St. Lawrence rift system comprises abundant large-scale normal faults displacing lower Paleozoic strata and underlying Grenville basement on the order of many 100's of meters along the Ottawa, Champlain, St. Lawrence, and Saguenay river valleys (Kumarapeli and Saull, 1966). These extensional faults generally cut discordantly across Grenville-aged structures instead of reactivating them. Mesoscopic-scale faulting of the lower Paleozoic rocks, with fault displacements ranging from less than a meter to several 10's of meters, has been recognized

locally throughout much of the Lake Ontario region outside of the St. Lawrence rift system (e.g., Liberty, 1960; Sanford, 1993; McFall, 1990, 1993). While it is well documented that the St. Lawrence rift system is associated with zones of elevated and persistent seismicity (Adams and Basham, 1991), patterns of earthquake activity in the Lake Ontario region have been characterized as diffuse, showing no persistent trends or apparent correlations with known or hypothesized geologic structures in the region (Stevens, 1994).

The Phanerozoic tectonic regime along the eastern margin of North America has been described in terms of two tectonic cycles, termed "Wilson cycles" (*see* Sanford and others, 1985). The major tectonic events affecting this region include: late Proterozoic-early Paleozoic rifting to form the Iapetus (proto-Atlantic) Ocean; middle to late-Paleozoic closure of the Iapetan ocean basin, accompanied by subduction of Iapetan crust and multiple arc-continent and continent-continent collisional episodes of Appalachian Mountain building; and Mesozoic rifting to form the present-day Atlantic Ocean (e.g., Rodgers, 1970; Bird and Dewey, 1970; Hatcher, 1972; Hatcher and others, 1989; Milkereit and others, 1992). The principal Phanerozoic tectonic elements of the southern Ontario region (shown on Figure 2-1) include (1) the Appalachian Orogen; (2) the Appalachian and Michigan basins; and (3) the basement-cored Findlay, Algonquin, and Frontenac arches.

It has been shown that, worldwide, the seismic potential of a stable continental region varies according to the degree of crustal extension that it experienced in the geologic past, and to a lesser extent, the age of the crust (Johnston and others, 1994). Three types of crust are identified in eastern North America (Figure 2-2): unrifted—the craton and the Appalachian fold belt; failed intracontinental rifts—the Ottawa and Saguenay aulacogens and the Reelfoot rift complex; and rifted passive continental margin—the Atlantic passive margin produced by the present opening of the Atlantic Ocean in late Mesozoic, and a relic passive margin produced by Iapetan rifting in late Proterozoic-early Paleozoic (Johnston and others, 1994). The north-northeast-trending faults along the Champlain and St. Lawrence river valleys, once attributed to a two-sided, failed intracontinental rift, are now recognized as part of the southeast-facing Iapetan margin (Wheeler, 1996a). The present-day Atlantic passive margin comprises transitional crust (continental-oceanic) and the extended and faulted inboard

continental shelf. Worldwide, Precambrian rifts such as the Grenville-age Midcontinent rift, do not, in general, appear to localize seismicity above background levels (Johnston and others, 1994).

Evidence of Iapetan rifting of the craton adjacent to the northern Appalachians is recorded within the St. Lawrence rift system (Kumarapeli and Saull, 1966) in the form of rift-related extensional structures, sediments, and magmatic/volcanic products that developed along the ancient continental margin of Laurentia (Kumarapeli, 1985). The rift structures include zones of en echelon faults parallel to the ancient margin, possible fracture zones transverse to the ancient margin, and two well-defined aulacogens (failed rifts), the Ottawa and Saguenay grabens. Syn-rift carbonatite complexes yielding K-Ar ages of approximately 565 Ma occur in both aulacogens, and a tholeiitic diabase dike swarm dated at 590 Ma is closely associated with the Ottawa graben (Kumarapeli, 1985, 1993).

The St. Lawrence rift system exhibits a branching, zigzag pattern, characteristic of other rift systems around the world (e.g., the Rhine graben and east African rift system) (Kumarapeli and Saull, 1966). The rift system extends northeastward from the junction of the Ottawa and Champlain valleys, down the St. Lawrence Valley, past the Saguenay graben (Kumarapeli and Saull, 1966), and along the sourthern coast of Labrador to the Labrador Sea (Gower and others, 1986) (Figure 2-3). The Sutton Mountains salient of the northern Appalachians coincides with the intersection of the St. Lawrence, Ottawa, and Champlain river valleys (Kumarapeli, 1985; Kumarapeli and St. Seymour, 1992). The Sutton Mountains have been interpreted as an ancient triple junction, or three-pronged rift, which formed as a prelude to the continental rifting of Laurentia and the opening of the Iapetan ocean basin (Kumarapeli and St. Seymour, 1992; Kumarapeli, 1993) (Figure 2-4). The flood basalts of the Tibbet Hill Formation (dated at 554 Ma) that occur in the Sutton Mountains are thought to represent the youngest flood basalts extruded at this triple junction (Kumarapeli, 1993). The approximately north-south-trending faults exposed along the margins of the Champlain Valley and the northeast-trending faults along the St. Lawrence Valley are believed to represent east-dipping and southeast-dipping listric faults that formed parallel to the rift arms that expanded to

oceanic dimensions. The east-west-trending Ottawa graben represents the failed arm of that rift system (Kumarapeli, 1985, 1993; Kumarapeli and St. Seymour, 1992).

Numerous authors, including Kumarapeli and Saull (1966), Woolard (1969), Martini and Bowlby (1991), and Adams and Basham (1989), have proposed that a midcontinent branch of the St. Lawrence rift system extends southwest through Lakes Ontario and Erie to the New Madrid rift complex (Braile and others, 1984; Hildenbrand and Hendricks, 1995) of the Mississippi Embayment (Figure 2-5). Adams and Basham (1991) subsequently retracted their speculative rift extension in notes added in proof in their 1991 publication. Evidence cited by Kumarapeli and Saull (1966) for the possible southwest extension of the St. Lawrence rift system includes: (1) an inferred continuous northeast-southwest-trending zone of seismicity extending from Quebec through the Findlay Arch, and into the Mississippi Embayment; (2) a local zone of normal faults in the axial region of the Findlay arch near Chatham, Ontario; (3) a northeast-southwest-trending zone of normal faults north of the Mississippi Embayment (the New Madrid rift) extending into the Wabash Valley; and (4) the occurrence of Cretaceous alkaline intrusive rocks in the St. Lawrence rift and along the margins of the Mississippi Embayment. Martini and Bowlby (1991) reported that the distribution and alignment of structures in the region and the inferred continuity of earthquakes extending from the St. Lawrence Valley through Lakes Ontario and Erie suggest a southwesterly, midcontinent extension of the rift system.

The Appalachian orogen lies approximately 400 km east of the Pickering and Darlington sites. Northern Appalachian orogenic events occurred from Ordovician to Permian time and consisted of several distinct tectonic episodes [*See* Stanley and Ratcliffe (1985) for a more detailed discussion]. Large-scale and deep-seated northwest-vergent thrust sheets and nappes composed of Grenville crystalline basement overrode the previously formed passive margin structure related to Iapetan rifting (Hatcher, 1972). The master detachment (floor thrust) underlying the mobilized thrust sheets is referred to as the Grenvillian ramp, or Appalachian detachment, which extends to a depth of about 30 km in the northern Appalachians (Hughes and Luetgert, 1991). The predominantly ductile deformation of the exhumed, older Paleozoic and Precambrian rocks within the internal portions (hinterland) of the Appalachian orogen is

associated with deformation at lower-crustal levels during Grenville and Appalachian orogenic episodes. These structures are locally overprinted by brittle deformation, which resulted from continental extension associated with the subsequent separation of North America and Africa during the Mesozoic (primarily Triassic - 245 to 208 Ma) episode of rifting. Rift basins that formed as a result of this extension are aligned with the Paleozoic contractional structures and locally involve the extensional reactivation of Paleozoic contractional faults (e.g., Lindholm, 1978; Swanson, 1986; Ratcliffe and others, 1986; Schlische, 1993).

The basic structural and tectonic framework of the passive continental margin of the Atlantic seaboard has been outlined by Klitgord and others (1988) (Figure 2-6). They subdivided the Atlantic margin into several crustal provinces (unrifted continental crust, rifted-continental crust, rift-stage crust beneath marginal sedimentary basins, marginal oceanic crust, and oceanic crust) based on the characteristic signatures seen in magnetic anomaly patterns (Klitgord and Behrendt, 1979), structural features imaged on seismic reflection lines (e.g., Klitgord and others, 1988), gravity models (e.g., Grow and others, 1979; Sheridan and others, 1988), geographically limited large-offset seismic data (e.g., Sheridan and others, 1979; LASE study group, 1986; Trehu and others, 1989a,b), and surface geologic data (e.g., Schlische, 1993). As discussed in the sections on individual seismic source zones, the key structural elements that mark the boundaries of these crustal provinces (e.g., the western limit of Mesozoic extensional structures) are used to define regional areal source zones that are characterized by similar crustal properties.

2.2 NEOTECTENIC SETTING

In the present tectonic setting the Pickering and Darlington sites lie within the stable continental region of the North American continent (Johnston and others, 1994). The geologically most recent, widespread, and unequivocal evidence for major tectonic activity in the region are Alleghanian (late Permian) thrust faults formed in the Appalachian foreland basin and late Triassic to late Jurassic normal faults along the Atlantic margin related to continental rifting and the subsequent opening of the Atlantic Ocean. However, historical seismicity along the St. Lawrence rift system, Charleston, South Carolina, and other concentrated zones; local geologic evidence of Cenozoic reactivation of faults; evidence of

paleoseismicity; and geologic and geodetic data indicative of regional and local crustal deformation are indicative of continuing neotectonic activity, albeit at much lower rates than during the last episode of major tectonic deformation.

Adams and Basham (1991) have reported that most large historical and instrumental earthquakes in eastern Canada have occurred near Paleozoic or younger rift zones. This is similar to stable continental region earthquakes worldwide (Johnston and others, 1994). The early Paleozoic St. Lawrence rift system, which is delineated by a persistent pattern of seismicity, is the postulated source of numerous large historical earthquakes in southeastern Canada (Adams and Basham, 1991). Seismicity along this rift systems is concentrated in a number of well-defined clusters, including the Ottawa River, Charlevoix, and lower St. Lawrence River seismic zones, which are all separated by relatively aseismic regions (Adams and others, 1995). Fault slip in in response to the current tectonic stress regime on pre-existing structures associated with the St. Lawrence rift system is believed to be the causal mechanism of this seismic activity (Adams and Bell, 1991; Anglin, 1984; Forsyth, 1981). Focal mechanisms from historical earthquakes within these seismic zones indicate mainly thrust events occurring at depths of 5 to 25 km within the Grenville basement (Adams and Basham, 1991).

Equivocal evidence for neotectonism has been found in the Lake Ontario region, and there are difficulties in distinguishing between deformation related to glacial processes and that related to deep-seated tectonic processes. Previously unrecognized lake-bottom features imaged by sidescan sonar in western Lake Ontario are interpreted by Thomas and others (1993) as potential evidence of neotectonism. These features include: (1) linear ridges, which are interpreted as stress-release pop-up structures in the underlying bedrock; (2) plumose structures, interpreted as the surface expression of stress release below the modern lake mud; and (3) dark lineations, which are thought to be sonar responses to intrusion of coarse sediment underlying the surface muds, or dewatering or degassing of sediment (Thomas and others, 1993). Lewis (1995), however, has attributed anthropogenic origins to both the plumose structures and dark lineations; the plumose structures representing anchor-drag marks, while the dark lineations represent steamship-ash debris trains (e.g., Cameron and

Lewis, 1994). The bedrock pop-ups were confirmed as neotectonic by visual observations from submersible diving in 1994 (Lewis, 1995).

In the Rochester basin of southeastern Lake Ontario, echosounding and shallow seismic surveys revealed features interpreted to be faults by Thomas and others (1993) and McQuest Marine (1995). The features, collectively referred to by Thomas and others (1993) as the South Ontario Structural Zone, have apparent offsets in the lake-bottom sediments of as much as 37.5 m and lie in a 15-km-wide zone along the southern margin of the postulated southwesterly extension of the St. Lawrence rift system (Thomas and others, 1993; McQuest Marine, 1995). Hutchinson and others (1993) and Lewis (1995) however, interpret the features as a buried drumlin field. These structures lie subparallel to the Hamilton-Presqu'ile fault zone, which extends northeast-southwest across Lake Ontario, parallel to the northern margin of the postulated St. Lawrence rift arm extension (Thomas and others, 1993). The east-northeast-west-southwest-trending structures of the South Ontario Structural Zone and the Hamilton-Presqu'ile fault zone, along with some of the features observed in western Lake Ontario, have been proposed as neotectonic evidence for the southwest continuation of the St. Lawrence rift system through Lakes Ontario and Erie (Thomas and others, 1993; McQuest Marine, 1995).

The postulated northwestern boundary of the late Proterozoic-early Paleozoic Iapetan rifted margin tectonic province lies approximately 80 km east of the Darlington site (Wheeler, 1995). This boundary (Figure 2-2) is interpreted as the cratonward limit of large normal faults in the Iapetan passive margin that have some long-term potential for reactivation in the present compressional tectonic regime (Wheeler, 1995). Wheeler (1995) has suggested that the Clarendon-Linden fault system (Figure 2-7) forms a segment of this western boundary. However, it is not known whether normal slip congruent with Iapetan rifting occurred on this structure during the late Proterozoic to Cambrian, and thus it cannot be defined unequivocally as an Iapetan structure (Wheeler, 1996a). There also is deep seismic evidence suggesting that the western boundary of the Iapetan margin may lie farther to the west, along the Central Metasedimentary Belt boundary zone (CMBBZ) of the Grenville Province (Milkereit and others, 1992) (Figure 2-7). However, the thickness (~43 km) and structure of Precambrian

crust in this region does not suggest significant Iapetan extension (Forsyth and others, 1994a,b; Zelt and others, 1994). Well-documented, high-angle Iapetan faults that cut Grenville basement are recognized along the eastern flank of the Adirondack Mountains (Kumarapeli, 1985); the western limit of these structures seems to be a more reasonable position for the western boundary of this province (P. Kumarapeli, written communication).

The eastern boundary of the Iapetan rifted margin is defined by a "zone of intense thinning" that separates highly attenuated Grenville crust to the southeast from extended but relatively intact crust to the northwest containing large Iapetan normal faults (Wheeler, 1994). Compressional reactivation of Iapetan normal faults is thought to play an important role in the seismic behavior of eastern North America (e.g., the Charlevoix, Giles County, and Eastern Tennessee seismic zones) (Adams and Basham, 1991; Anglin, 1984; Bollinger and Wheeler, 1983; Johnston and others, 1985; Teague and others, 1986; Wheeler and Johnston, 1992; Powell and others, 1994; Wheeler, 1995).

West of the Iapetan rifted margin lies the Precambrian Central Craton (Wheeler and Johnston, 1992), which is composed of relatively flat-lying Paleozoic platform sediments overlying non-extended basement crust of the Grenville (~1100 Ma) and Eastern Granite-Rhyolite Provinces (1400-1500 Ma) (Bickford and others, 1986) (Figures 2-2 and 2-8). The rate of historical seismic activity in this region is low and appears typical of stable cratonic crust (Johnston and others, 1994). In general, seismic activity and the geologic conditions most associated with earthquake activity in the stable continental region of central and eastern North America increase toward the east, away from the Precambrian Central Craton and toward the rifted passive continental margin.

2.3 CONTEMPORARY STRESS FIELD

Southeastern Canada is included in a broad midplate compressive stress province that includes most of the central and eastern United States and possibly much of the western North Atlantic basin to within about 250 km of the Mid-Atlantic ridge (Zoback and others, 1986). The stress field throughout this region is characterized by a nearly horizontal, northeast- through east-striking axis of maximum compressive stress (Sbar and Sykes, 1973; Hermann, 1979; Zoback

and Zoback, 1991) (Figure 2-9). The overall uniformity in the direction and relative magnitude of the midplate stress pattern suggests a far-field tectonic source, and ridge-push (due to accretion of oceanic crust along the mid-Atlantic ridge) is preferred over basal drag as the primary source of stress in the midplate region (Zoback and Zoback, 1991).

Most earthquakes in this midplate stress province have strike-slip, oblique-reverse, or reverse fault mechanisms (Dewey and others, 1991). Well-constrained focal mechanisms for earthquakes that have occurred in eastern North America in the past 25 years indicate primarily thrust (reverse) fault and/or strike-slip events (Yang and Aggarwal, 1981; Zoback, 1992; Adams and Bell, 1991; Hermann, 1978; Seeber and Armbruster, 1988). Northwest- and northeast-trending faults are most favorably oriented for reactivation by reverse and/or strike-slip displacement according to recent interpretations of the orientations of the contemporary regional stress components (Zoback, 1992; Adams and Bell, 1991). Reverse faulting on planes striking other than northeast and northwest, however, is not precluded (Adams, 1995). The available data from the northeastern United States generally indicate high horizontal stresses at shallow depths (less than about 0.5 km).

Focal mechanisms for most earthquakes in southeastern Canada and the adjacent United States indicate that thrust and/or strike-slip mechanisms dominate the region, consistent with the regional field (Adams and Bell, 1991). Equivocal evidence from within the seismogenic crust suggests that stress orientations may change with depth: shallower earthquakes (<10 km) tend to be more consistent with ENE compression, whereas deeper events tend to be less consistent (Adams and Bell, 1991). Some stress indicators from the region show anomalous directions; however, it is uncertain if these are real, second-order features of the regional stress field (Adams, 1995). The available data from eastern Canada indicate that deviatoric stresses throughout the shield generally exceed those necessary to generate earthquakes (Adams, 1995). However, there is no evidence to suggest that particular regions are more likely to generate earthquakes based on the stress indicators (Adams, 1995).

Stress indicators in the Lake Ontario area are in general agreement with the more regional datasets referred to above. A systematic set of neotectonic[1] joints along the Niagara escarpment suggests a N60E orientation for the maximum horizontal compressive stress (Gross and Engelder, 1991). This neotectonic, ENE-trending joint set is also observed regionally throughout the northeastern United States (Engelder, 1982). The orientations of post-glacial pop-up structures and quarry-floor buckles measured in southern Ontario also imply an ENE orientation of the maximum horizontal compressive stress in the near surface (Wallach and others, 1993). Stress orientations measured in a 300-m-deep borehole at the Darlington site indicate that within the Paleozoic rock sequence, the orientation of the maximum horizontal compressive stress ranges from N70E at a depth of 75 m to N32E at depth of 207 m (Haimson and Lee, 1980). In the underlying Grenville basement, an average orientation of N23E was measured to a depth of 300 m (Haimson and Lee, 1980).

2.4 Local Tectonic and Geologic Setting

The Pickering and Darlington sites are located on the northern shore of Lake Ontario. Lake Ontario is situated along the northwest margin of the Appalachian sedimentary basin (Figure 2-1), which is bounded by Paleozoic and Precambrian bedrock highs: the Algonquin arch to the west; the Frontenac arch to the north; and the Adirondack Mountains to the east. The Algonquin and Findlay arches to the west separate the Paleozoic Appalachian and Michigan basins.

Bedrock underlying the Lake Ontario region consists of gently dipping sequences of early Paleozoic sandstones, limestones, and shales, which unconformably overlie exhumed Mesoproterozoic mid-crustal crystalline rocks of the Grenville Province (Figure 2-10). Grenville-age rocks are exposed in the Canadian Shield, approximately 50-100 km to the north of Lake Ontario. The overlying Paleozoic sequence thickens and becomes younger to the south, toward the axis of the Appalachian basin. Borehole and seismic data indicate the

[1] Neotectonic is herein defined as being formed within the present-day, compressional tectonic stress regime, which was established roughly in the early to mid Cenozoic subsequent to Mesozoic rifting of the present-day North Atlantic Ocean.

Paleozoic cover is about 200-300 m thick beneath western Lake Ontario, increasing to about 10 km thick in New York and Pennsylvania (Haimson and Lee, 1980; Pohn and Coleman, 1991).

Regional joint and fracture systems have been well documented in the lower Paleozoic rocks exposed around Lake Ontario. Numerous joint sets, with trends ranging from 005° to 135° are recognized on the Niagara Peninsula and in upper New York State (e.g., Williams and others, 1985; Gross and others, 1992; Gross and Engelder, 1991). In the Head Lake-Burliegh Falls area of south-central Ontario, the dominant joint sets trend 027°, 091°, and 152° (Rutty and Cruden, 1993). Along the northeast margin of Lake Ontario, joint trends ranging from 015° to 130° are recognized (McFall and Allam, 1991), and in the western Lake Ontario area, the dominant joint trends are 20°, 70°, and 135° (Rogojina and others, 1995). Gross and Engelder (1992) have suggested that east-northeast-trending (060-070°) joints measured in the region are neotectonic structures, because they parallel the orientation of the regional maximum horizontal compressive stress. The approximately east-west-trending joints and the northeast- and northwest-trending joints measured in the region are attributed to systems that developed during the various Appalachian orogenic episodes (Williams and others, 1985; Gross and others, 1992; Engelder and Geiser, 1980; Rutty and Cruden, 1993). Sanford and others (1985) have proposed that a regional-scale network of east-west-trending and northeast-southwest-trending fractures developed in southwestern Ontario during Iapetan rifting. They further suggested that subsequent compressional and extensional tectonic regimes during the Phanerozoic periodically reactivated these fractures through uplift of the basement-cored arches marginal to the Paleozoic sedimentary basins, and small-scale differential vertical displacement across the fracture/fault-bounded blocks.

Local occurrences of mesoscopic-scale faults that offset the lower Paleozoic cover rocks have been documented throughout the Lake Ontario region (e.g., Liberty, 1960; McFall, 1993; Sanford, 1993; Grier, 1995). Sanford (1993) documented numerous small high-angle normal faults, with vertical offsets generally of less than 1 meter in the Paleozoic rocks near the Paleozoic-Precambrian contact north of Lake Ontario. The maximum vertical offset for a single fault observed by Sanford (1993) was about 30 m. In Prince Edward County, McFall

(1993) documented numerous faults with vertical offsets of the lower Paleozoic rocks ranging from less than 1 meter to about 40 meters. One east-west-trending strike-slip fault observed by McFall (1990) cuts a Jurassic-aged ultramafic dike (Barnett and others, 1984). The amount of displacement on this post-Jurassic fault is unknown, however, due to the lack of exposed marker horizons (McFall, 1990).

Northeast-southwest-trending physiographic lineaments have been documented in both the Precambrian and Paleozoic rocks to the north and east of Lake Ontario (Lowman and others, 1987; Singhroy, 1988), along the postulated southwest projection of the St. Lawrence rift system in the upper St. Lawrence Valley. On the Frontenac arch in northern New York State, Gauthier and Benn (1996) identified two prominent northeast-striking faults that juxtapose Paleozoic strata on the northwest against Grenville basement on the southeast. These faults are associated with brittle deformation and are interpreted to be reactivated ductile Grenville structures (Gauthier and Benn, 1996). Williams (1991) documented as much as 1000 m of cumulative displacement on numerous small-displacement normal faults related to the Ottawa graben in southeastern Ontario. In addition, Grier (1995) documented brittle faulting on a series of northeast-trending faults in the upper St. Lawrence Valley, between Kingston and Cornwall. In recent years a number of anomalous features have been observed on the bottom of Lake Ontario (e.g., Thomas and others, 1993), which have been interpreted to support the southwest extension of the St. Lawrence rift system as far south as Hamilton, Ontario. The origin of these features is equivocal; some have been interpreted as indicators of neotectonic activity.

Late Cenozoic crustal deformation related to glacial processes is recognized in the southern Ontario region. The Laurentide ice sheet is estimated to have been as much as 3000 m thick during the last glacial maximum, and apparently was associated with depression and rebound of the shield on the order of 100's of meters (Bloom, 1991). Differential uplift and tilting of the entire Lake Ontario basin has been attributed to post-glacial isostatic rebound related to the removal of the Laurentide ice sheet (Martini and Bowlby, 1991). Coleman (1937) documented tilting of the glacial Lake Iroquois shoreline, with the northeastern margin uplifted relative to the southwestern margin. This differential uplift of the land has been

shown to be continuing at a rate of about 0.01 ft/yr in the northeast part of the Lake Ontario basin relative to that in the southwest (Clark and Persoage, 1970). The Laurentide ice sheet was the last of several great ice sheets to cover parts of Ontario during the Quaternary Period (Barnett, 1992). It is not known if the cumulative effects of these great ice sheets have contributed to the small-scale deformation of the Paleozoic strata observed in the Lake Ontario region.

Bedrock pop-up structures are elongate anticlinal features that form at the bedrock surface by buckling under horizontal compression. They have been documented throughout southern Ontario, and most can be shown unequivocally to be post-glacial, and thus neotectonic (e.g., Wallach, 1990; Wallach and others, 1993; Rutty and Cruden, 1993; McFall, 1993). Pop-ups generally appear as narrow ridges with open fractures located along their crests, and some exhibit an en echelon geometry (Rutty and Cruden, 1993). The formation of pop-ups in this region can be attributed in most cases to the retreat of the Laurentide ice sheet and the reduction of the vertical component of crustal stress. Pop-ups generally form perpendicular to the direction of the local maximum horizontal compressive stress, and thus can be used as local stress indicators (Wallach and others, 1993). In some cases, pre-existing joints and fractures have controlled pop-up orientations (Rutty and Cruden, 1993). The majority of pop-ups documented in the Lake Ontario region trend northwest, roughly orthogonal to the regional ENE maximum horizontal compressive stress (Wallach and others, 1993). Sanford (1995) noted that northeast-trending pop-ups are dominant structural features on the bottom of Lake Ontario.

Numerous deep-seated Precambrian (Grenville) ductile structures, expressed as well-defined aeromagnetic anomalies, occur beneath the Paleozoic cover rocks underlying the Lake Ontario region (Figures 2-7 and 2-11). Although deep seismic data collected across some of these structures in Lakes Erie and Ontario suggest that they have been tectonically stable since early Paleozoic times (e.g., Forsyth and others, 1994a,b; Milkereit and others, 1992), they have been identified as potentially active features based on local outcrop-scale brittle faults and fractures observed in the Precambrian and overlying Paleozoic rocks in their vicinity (Wallach and Mohajer, 1990). Eyles and others (1993) have presented evidence suggesting that relict

drainage systems that cut across the lower Paleozoic rocks in the western Lake Ontario region have been controlled in part by reactivation of some of the Precambrian structures associated with these geophysical lineaments.

The most prominent of these features is the north-northeast-trending Niagara-Pickering linear zone, which is defined by a conspicuous linear magnetic anomaly and a less prominent gravity anomaly in the Grenville basement, and physiographic features in the overlying Paleozoic cover (Wallach and Mohajer, 1990; Wallach, 1990). This feature, which extends beneath the western end of Lake Ontario, coincides with the CMBBZ where it is exposed within the Canadian Shield to the north (Figures 2-7, 2-8, and 2-11). The CMBBZ is a major terrane boundary separating rocks of the Central Metasedimentary belt from those of the Central Gneiss belt in the Grenville Province (Forsyth, 1981; Forsyth and others, 1994). The Niagara-Pickering linear zone parallels the Clarendon-Linden fault system (Fakundiny and others, 1978) of upper New York State, which is inferred to extend northeastward beneath Lake Ontario (Hutchinson and others, 1979) and into the Canadian Shield.

A second prominent magnetic lineament in the Grenville basement extends from Burlington to Toronto, parallel to the north-northeast trend of the Lake Ontario coastline (Figures 2-7 and 2-11). Thomas and others (1993) interpreted bedrock pop-up structures and linear patterns along this feature on the floor of Lake Ontario as evidence of neotectonic deformation. Mohajer (1993) suggested that this lineament and a parallel linear magnetic anomaly to the west form a 20-km-wide zone of concentrated seismicity, termed by Mohajer (1993) the Toronto-Hamilton seismic zone.

A third feature, the Georgian Bay linear zone, is delineated on the basis of linear, discontinuous magnetic and gravity anomalies in the Grenville basement (Wallach and Mohajer, 1990) (Figure 2-12). This northwest-trending feature is interpreted to extend along the eastern coastline of Georgian Bay, through Lake Ontario to Attica, New York, and is believed to correspond to a prominent basement structure in the region that has been hypothesized to be seismically active (Wallach and Mohajer, 1990).

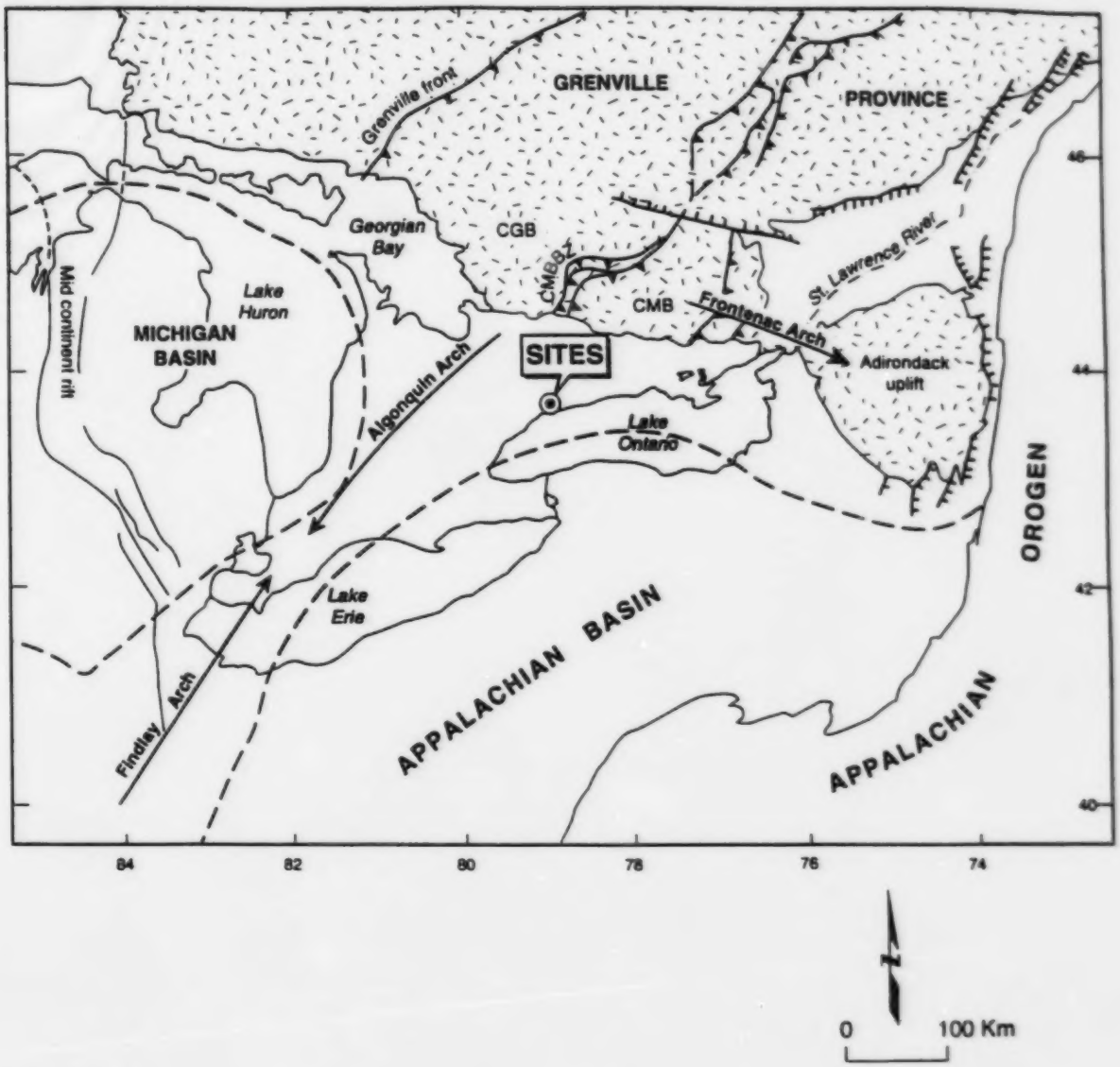


Figure 2-1 Simplified tectonic map of the eastern Great Lakes region (modified from Muehlberger, 1991)

Figure 2-2 The three types of seismogenic continental crust of eastern North America. COB is continental/oceanic crust boundary (modified from Johnston, 1995)

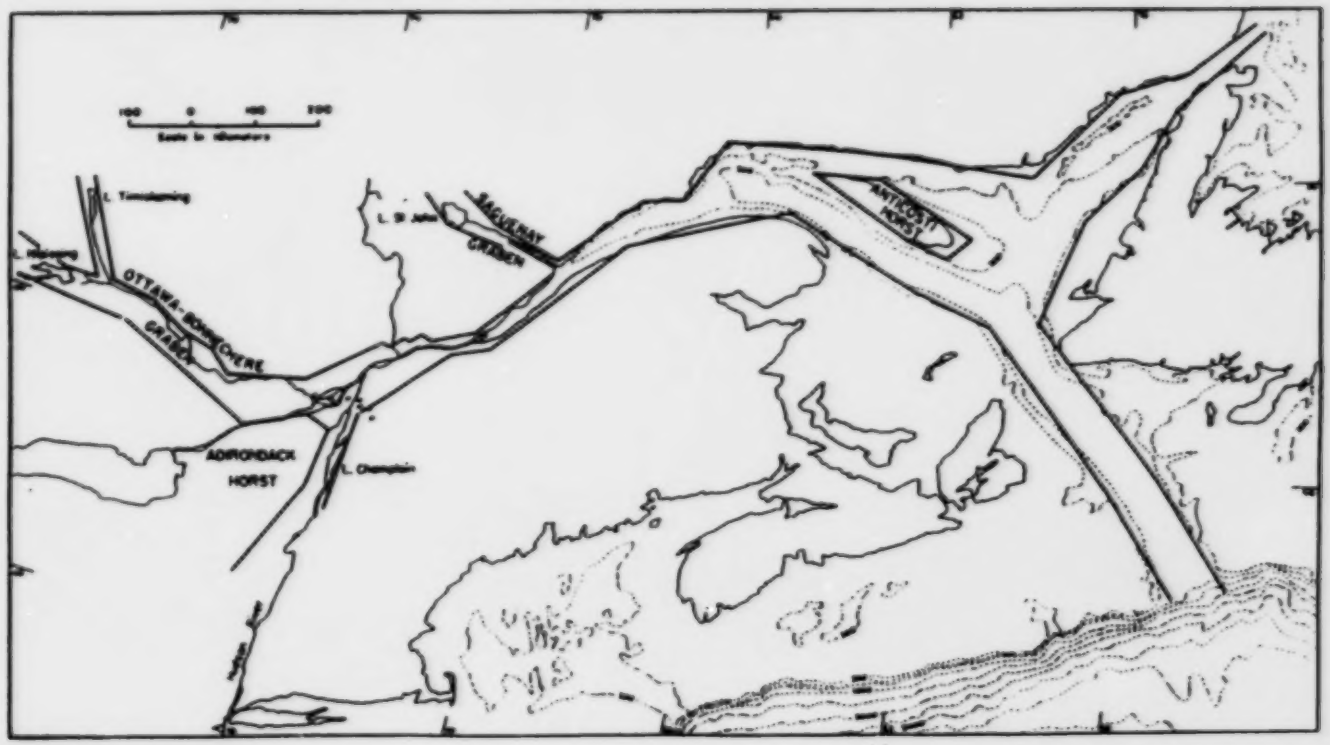


Figure 2-3 Rift zones of the St. Lawrence system (modified from Kumarapeli and Saull, 1966)

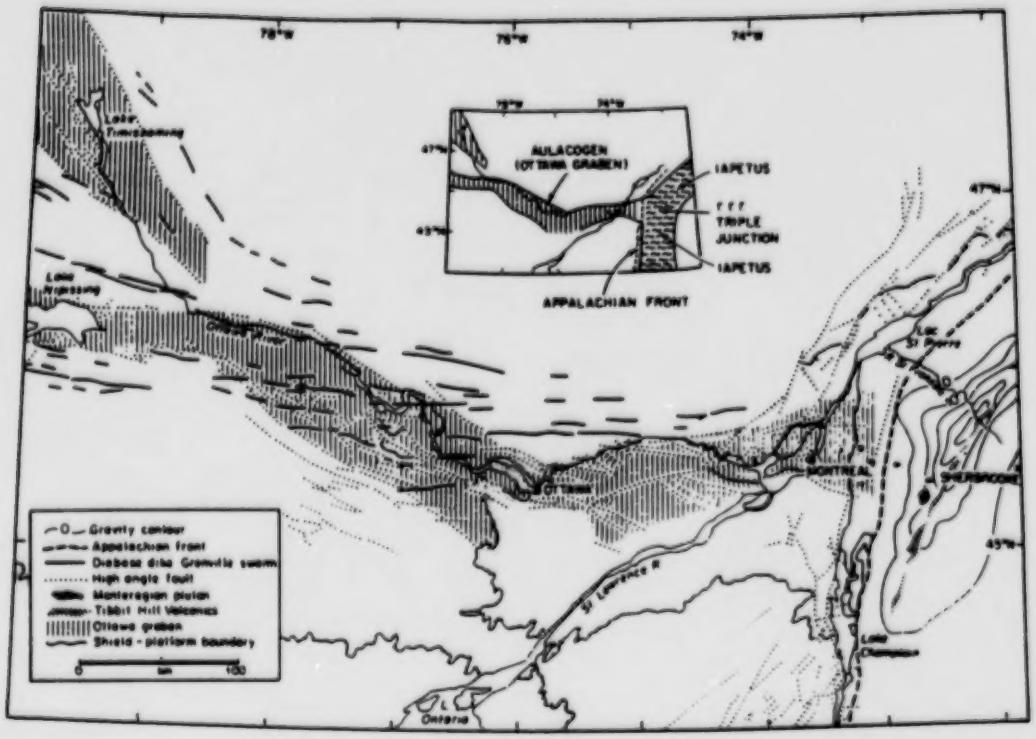


Figure 2-4 Geological elements of the Ottawa graben. Inset, interpretive sketch of the postulated Sutton Mountains *rrr* triple junction (from Kumarapeli and others, 1981)

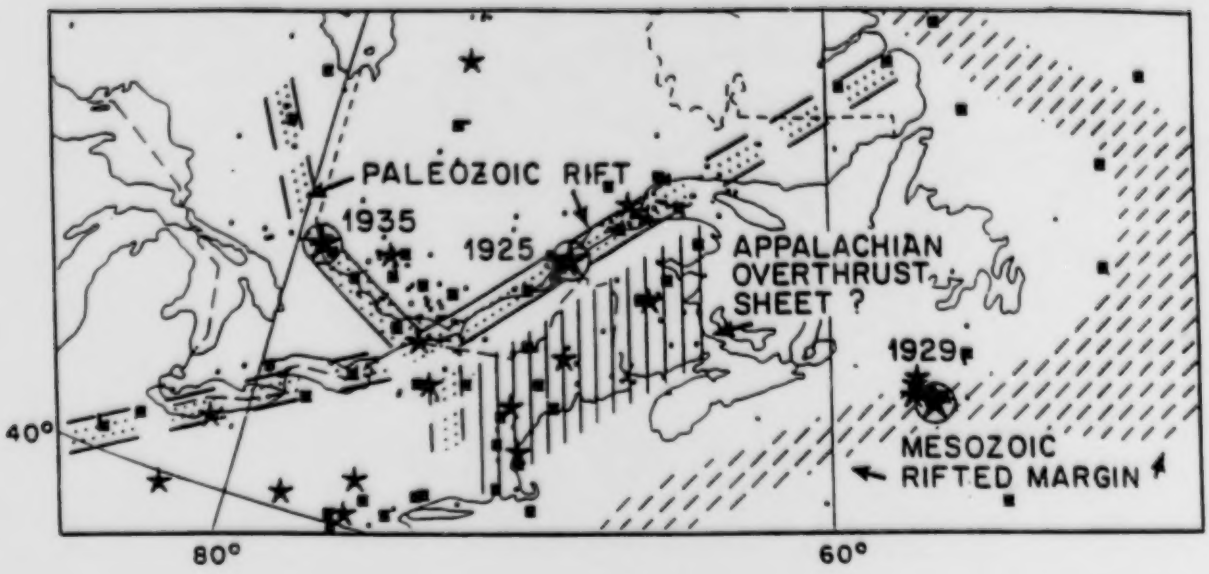


Figure 2-5 St. Lawrence rift showing postulated SW midcontinent extension (from Adams and Basham, 1989)

Figure 2-6 Map showing tectonic and geophysical features along part of the U.S. Atlantic continental margin (from Klitgord and others, 1988)

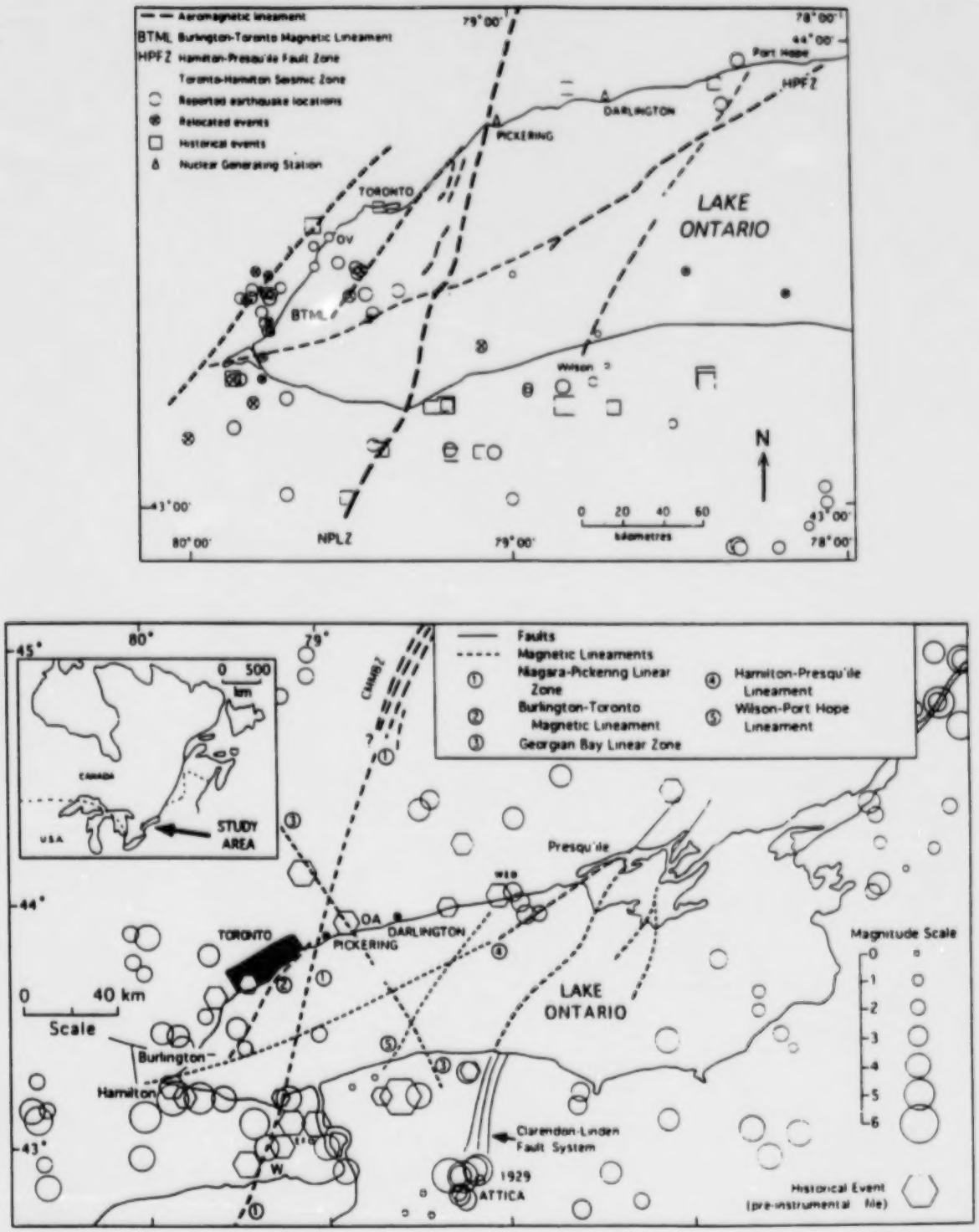


Figure 2-7 Seismicity and potentially seismotectonic features of Lake Ontario (from Mohajer, 1993)

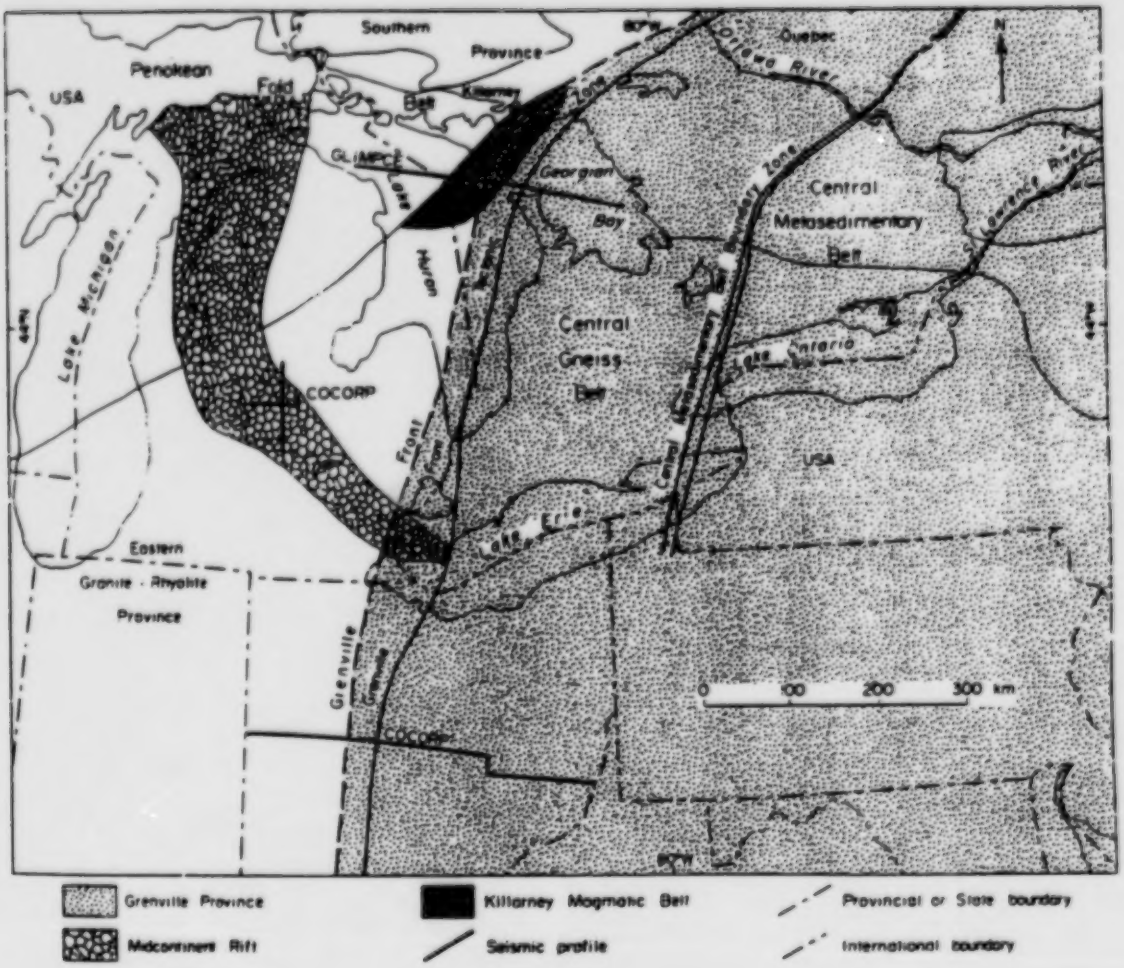


Figure 2-8 Principal subdivisions of Precambrian rocks in the Great Lakes region (from Easton, 1992)

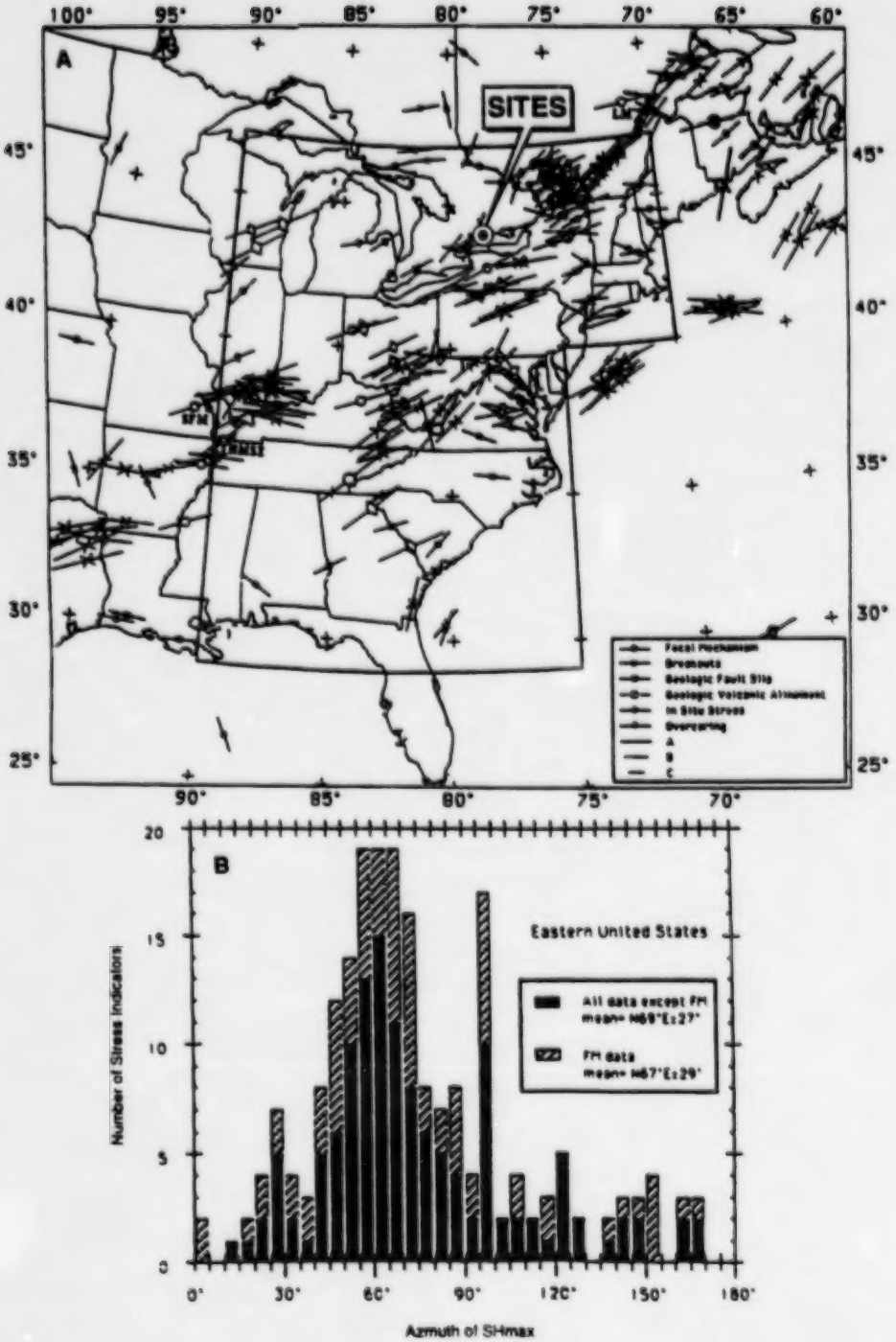


Figure 2-9 Stress map of central and eastern North America showing directions of maximum horizontal compressive stress. Data points within polygon are shown in the histogram (from Zoback and Zoback, 1991)

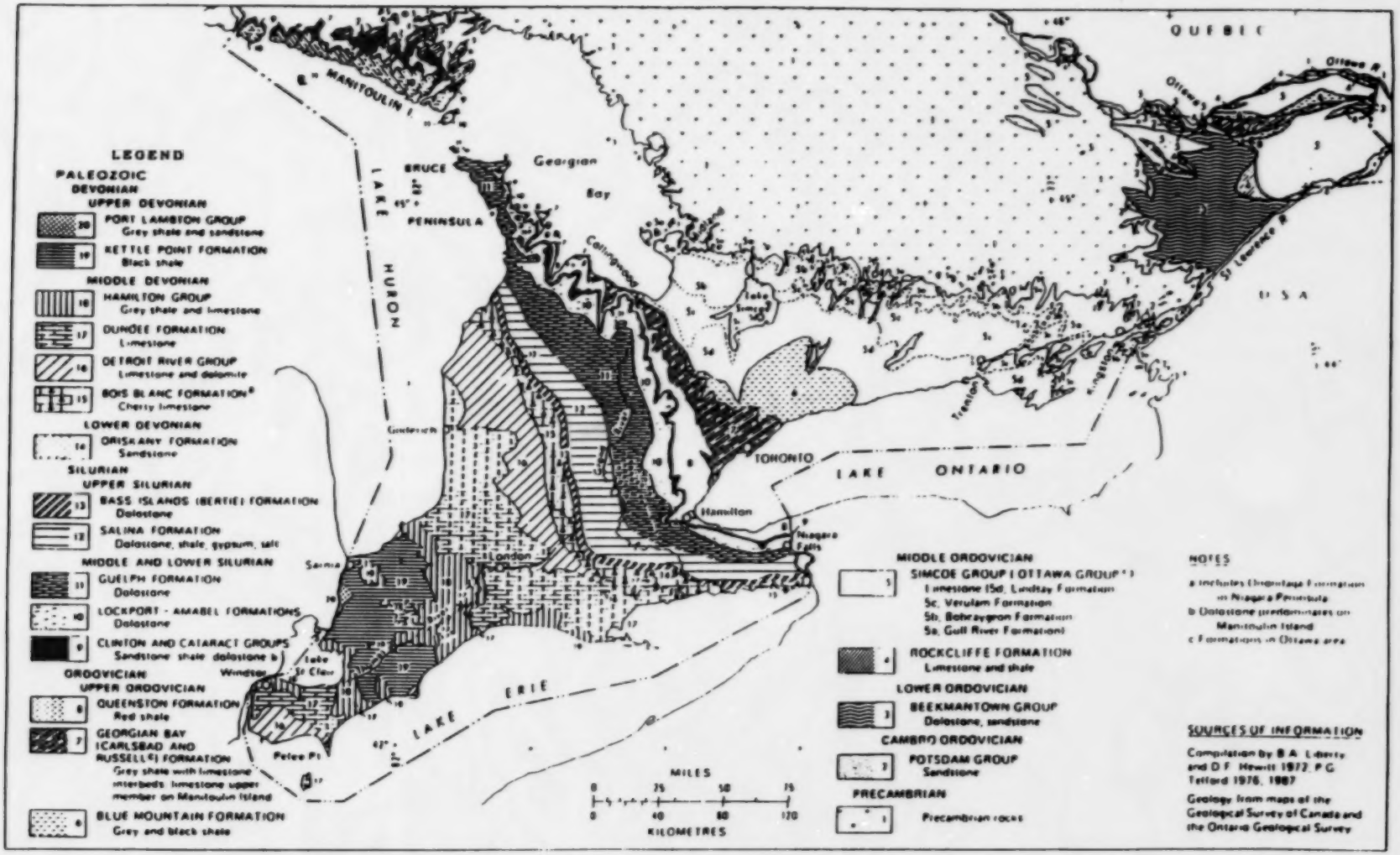


Figure 2-10 Geologic map and stratigraphic units of southern Ontario (from Martini and Bowlby, 1991)

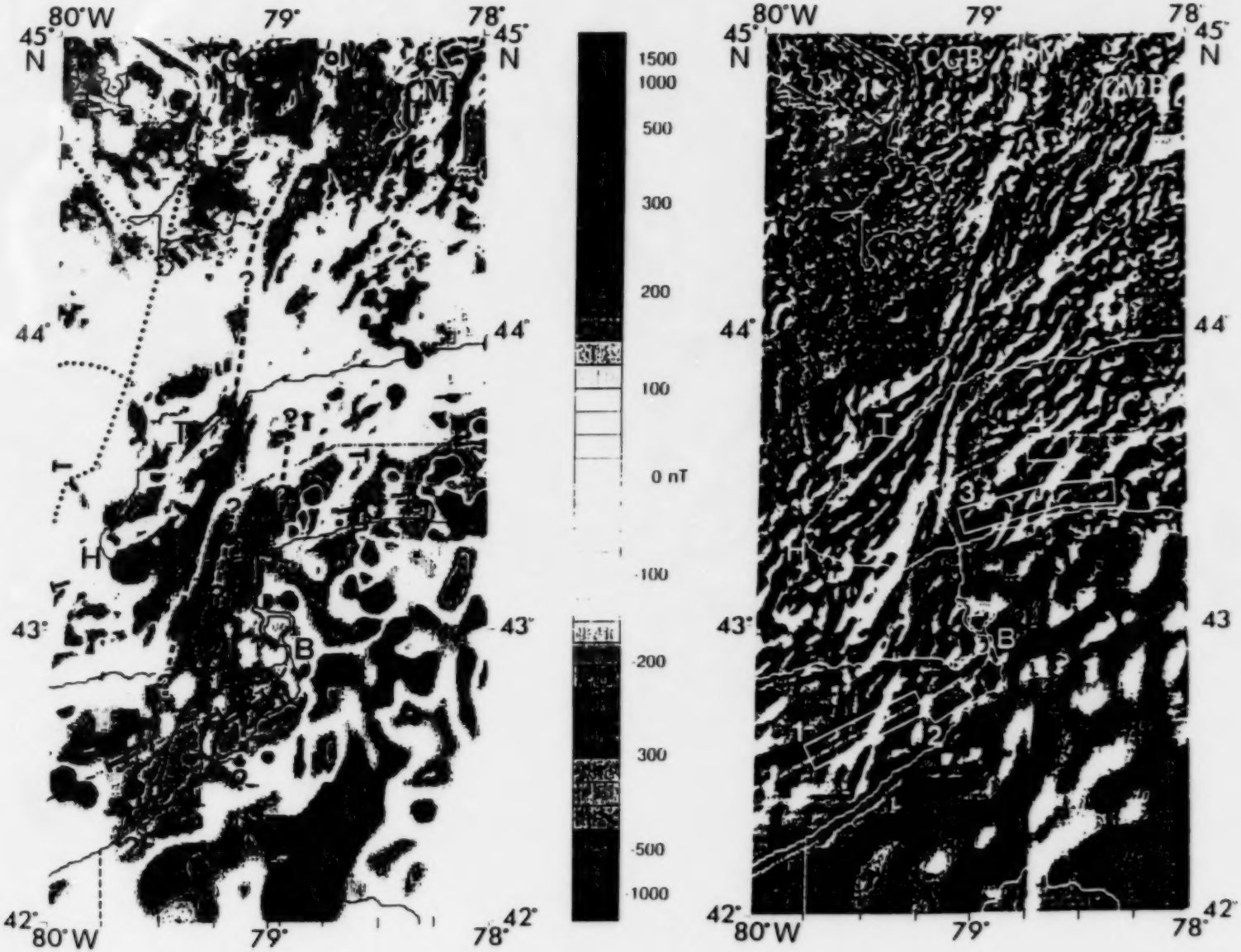


Figure 2-11 Residual total magnetic field and shaded relief image illuminated from an apparent sun to the southeast. Dashed lines indicate magnetic trends coincident with interpreted boundary zone of the CMBBZ. Dotted lines are Grenville subdivisions; fine dotted line shows northern limit of Paleozoic cover. Rectangles 1, 2, 3, and 4 are locations of seismic profiles. B, Buffalo; M, Minden; T, Toronto; H, Hamilton (from Forsyth and others, 1994a)

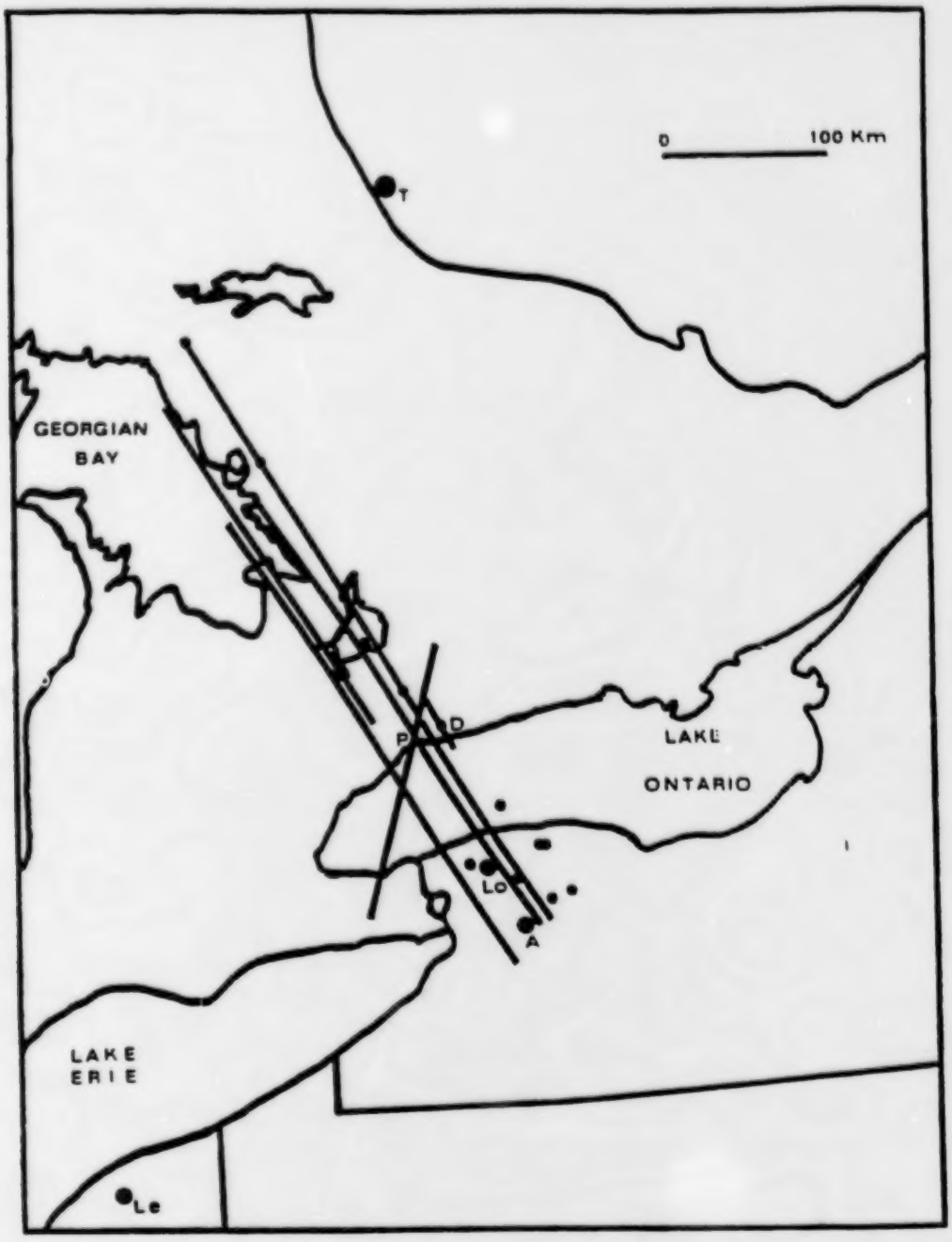


Figure 2-12 Structural zones in western Lake Ontario. Line oriented NE-SW is the Niagara-Pickering linear zone; lines oriented NW-SE comprise the Georgian Bay linear zone. Dots are earthquake epicenters; A, Attica; Le, Leroy; Lo, Locport, and T, Timisksming. P and D are the Pickering and Darlington Nuclear Generating Stations, respectively (from Wallach, 1990)

3.0 SEISMIC SOURCES

Following standard practice in PSHA, *seismic sources* are identified that define the future locations of earthquakes (*See* general discussion of seismic sources in Section 1). The boundaries of seismic sources are usually drawn based on geologic structures and features, as well the spatial pattern of seismicity. Seismic sources are distinguished from each other usually on the basis of judged differences in maximum magnitude and/or recurrence rate. These sources are usually expressed as areal source zones. Further, some seismic sources represent particular tectonic features (faults or zones) that are judged to have some potential for being seismogenic.

In this study, we have focused considerable attention on capturing particular *tectonic hypotheses* in constructing seismic source interpretations. That is, the spatial configuration of alternative seismic sources is based in most cases on a particular tectonic hypothesis that has been proposed. Examples include the possible southwestern midcontinent extension of the St. Lawrence rift system, a potentially seismogenic Niagara-Pickering linear zone, a potentially seismogenic Toronto-Hamilton seismic zone, and a potentially seismogenic Rouge River fault zone. In order to include these hypotheses in the analysis, it is important to also consider the uncertainty that exists regarding their *seismogenic potential* (seismogenic potential is defined here as their potential for generating $M \geq 5$ earthquakes in the present tectonic regime).

In addition to the classical approach to the definition of seismic sources, which includes considerations of tectonics and tectonic features, we also provide for the use of observed seismicity alone to express the future locations of earthquake occurrence. We do this by either defining seismic source zones enclosing regions of elevated seismicity or by "smoothing" the spatial pattern of observed earthquakes (or the local rate of observed earthquakes). An advantage is that the smoothing approach does not require drawing source boundaries and there is no need to assume a uniform rate of recurrence within a source boundary (*See* later discussion in Section 5 on recurrence methodologies). Both approaches, however, are based on an assumption of a certain level of 'spatial stationarity, that is, the locations of past (observed) earthquakes are indicators of the locations of future seismicity.

Seismic sources identified for this study consist of two types: *regional seismic sources* and *local seismic sources*. Regional sources are all areal source zones and are located within about 300 km from the Pickering and Darlington Nuclear Generating Stations (in some cases, more distant sources are also included). Local seismic sources exist within about 100 km of the sites and are intended to represent the possibility that earthquakes are localized within the regional source zones. The local zones are defined based on various tectonic hypotheses regarding certain interpreted tectonic features. Because the earthquake potential of these features is uncertain, the probability of each feature being seismogenic is assessed explicitly. "Seismogenic potential" is defined for this study as having the potential of generating magnitude $m_b \geq 5$ earthquakes. All of the regional seismic sources are assumed to be seismogenic (probability [seismogenic] = 1.0). The local sources have assessed probabilities of being seismogenic. In the case where a particular local source is assessed to be seismogenic, it will localize earthquakes in the hazard analysis relative to the regional source zone within which it lies; if the local source is not seismogenic, earthquakes will be modeled in the hazard analysis as occurring within the regional source zone.

Below, interpretations of regional seismic sources are presented first, followed by interpretations of local seismic sources.

3.1 REGIONAL SEISMIC SOURCES

The regional seismic source zones are shown on Plates 1 through 14. Regional geological source zones are represented on Plates 1 through 10, and a set of geological/seismological composite source zones from Atkinson (1990) are shown on Plate 11. Regional seismicity is shown on Plate 12, and the regional and local seismicity zones of Adams and others (1996) are shown on Plates 13.

3.1.1 Iapetan Rifted Margin (IRM) Source Zone

The Iapetan Rifted Margin source zone (IRM) extends northeast-southwest across the entire study area and encompasses a terrane of known and inferred normal faults that formed parallel to the passive margin of Laurentia during the late Proterozoic-early Paleozoic opening of the

Iapetus Ocean (e.g., Wheeler, 1995, 1996b) (Plates 1, 2, and 3). Compressional reactivation of favorably oriented Iapetan faults has been postulated as the causal mechanism for several seismically active regions in eastern North America, including Charlevoix and the lower St. Lawrence Valley, Quebec (Adams and Basham, 1991); Giles County, Virginia (Bollinger and Wheeler, 1988; and eastern Tennessee (Powell and others, 1994). Adams and others (1994) considered the 1925 Charlevoix event (M 6.2, Bent (1992)) as the paradigm earthquake for the IRM source zone, however, they acknowledged that the great depth extent of seismicity and the length of the faults suggest much larger earthquakes are possible.

Bollinger and Wheeler (1988) suggested that the southeastern boundary of the Iapetan rifted margin coincides approximately with a steep eastward rise in the unfiltered Bouguer anomaly field. This gravity rise, referred to as the Appalachian (Piedmont) gravity gradient, is interpreted to represent the transition from thick continental to less thick, and possibly more mafic (transitional) crust to the east (James and others, 1968; Kane and Long, 1981; Hatcher and Zietz, 1980; Hutchinson and others, 1983). Bollinger and Wheeler (1988) and Wheeler (1996b) suggested that most Iapetan faults occur in the relatively intact continental (Grenville-age) crust of the North America, west of the gravity rise.

Based on analysis of deep seismic reflection profiles across parts of the Appalachians and the Coastal Plains, Wheeler (1996b) inferred the southeast boundary of preserved Iapetan normal faults to coincide with a narrow zone of intense (southeastward) thinning of Grenville crust beneath the Appalachian thrust sheets. The zone of intense thinning extends along the Appalachians and is approximately coincident with the Appalachian gravity gradient. Wheeler (1996b) noted that in the zone of intense thinning and farther southeast, the profiles show structures that disrupted or destroyed the Grenville crust and any Iapetan faults within it during Paleozoic compressional and Mesozoic extensional deformation. The Iapetan faults likely decrease in size, abundance, and slip, gradually and irregularly northwestward into the North American craton over a distance of perhaps 100-200 km (Bollinger and Wheeler, 1988).

Wheeler (1995) reported that the northwestern boundary of the Iapetan rifted margin is defined by the locations of known and inferred Iapetan normal faults, that are currently both seismic

and aseismic. The northwestern boundary of the Iapetan rifted margin also generally coincides with the transition from a more seismically active continental rim to the east, to a relatively less active cratonic interior to the west. The Clarendon-Linden fault system has been postulated to form a segment of this western boundary; however, it is not known to be a true Iapetan structure (Wheeler, 1996a) (Figures 3-1 and 2-2). Well-documented Iapetan faults are recognized along the eastern flank of the Adirondack Mountains, well over 100 km east of the Clarendon-Linden fault system (Kumarapeli, 1993). Wheeler (1995) noted that the majority of Iapetan faults that constrain the northwestern boundary occur as groups bounding large graben, or as closely spaced steps that drop prefaulting basement down to the southeast. The majority of Iapetan structures used by Wheeler (1995) to define this boundary posses evidence of late Proterozoic, as well as early Paleozoic (Cambrian) slip. Wheeler (1995) noted that a few unrecognized Iapetan faults may occur northwest of this boundary, however, he suggested their effect on seismic hazard probably is not significant.

Milkereit and others (1992) have suggested that the CMBBZ, which lies approximately 100 km to the west of the Clarendon-Linden fault system (Figure 3-1), displays characteristics of an Iapetan structure. A middle to late Proterozoic half graben imaged from deep seismic reflection data collected across the CMBBZ in eastern Lake Erie was interpreted to be the result of either pre-Appalachian (i.e., Iapetan) continental rifting, or terminal collapse of the Grenville orogen (Figure 3-2). This feature involved extensional reactivation of a gently east-dipping Grenville-age thrust fault in the CMBBZ, as opposed to cutting discordantly through the gently east-dipping boundary zone (Milkereit and others, 1992). The seismic data indicate that the lower Paleozoic sediments overlying this half graben are not significantly deformed, at the data resolution scale (Milkereit and others, 1992).

Based on the foregoing information, we consider three alternatives for the position of the northwestern boundary of the IRM source zone: the regional Iapetan rifted margin boundary as substantially defined by Wheeler (1995) (Plate 1a); the CMBBZ as suggested by Milkereit and others (1992) (Plate 2a); and the western limit of known Iapetan normal faults in the Adirondack uplift (Plate 3a). A low weight (0.1) is assigned to the CMBBZ representing the northwest boundary of Iapetan rifting based on: (1) the absence of large-scale, high-angle,

down-to-the-east step faults cutting through the gently east-dipping Grenville reflectors in the seismic data, (2) the thickness (~ 43 km) of the Grenville crust adjacent to the CMBBZ (Zelt and others, 1994; Forsyth and others, 1994a), (3) the well-documented occurrences of post-orogenic extension due to collapse in the Grenville Orogen within the Canadian Shield (e.g., van der Pluijm and Carlson, 1989; Busch and others, 1996), and (4) the absence of significant early Paleozoic normal slip overlying the half graben imaged in Lake Erie. We assign a weight of 0.5 to the boundary lying along the western flank of the Adirondack uplift, and a weight of 0.4 to the boundary as defined by Wheeler (1995). The western Adirondack boundary is given slightly higher weight because it marks the limit of known Iapetan faults in the Lake Ontario region, and the Clarendon-Linden fault system cannot be shown unequivocally to be an Iapetan structure. Uncertainty in the location of this boundary could be reduced by examination of the thickness of Cambrian units adjacent to the Clarendon-Linden fault system versus the CMBBZ.

In view of the uncertainty associated with the location of the western boundary of the IRM source zone, we make an assessment of the significance of the IRM/SGR source zone boundary. Johnston and others (1994) noted that in stable continental regions worldwide, Precambrian rifts, such as the Midcontinent rift system, are considered to be incorporated into the craton and are aseismic to background seismicity levels. Iapetan rifting occurred from very late Proterozoic and continued into the earliest Paleozoic (Johnston and others, 1994). Thus, Iapetan rifted crust is considered a Paleozoic domain by Johnston and others (1994). The largest earthquake to occur in the Iapetan rifted margin is the 1925 M 6.2 Charlevoix event, though the 1663 event (M~6.6) may have been larger (Bent, 1992). Other notable events to occur in Iapetan rifted crust are the 1935 **M** 6.2 Timiskaming earthquake, and the 1988 **M** 5.9 Saguenay event. The largest known earthquake to occur in the unrifted crust of eastern North America is the 1989 M 6.0 Ungava event, which is considered one of the rarest and least explainable types of earthquakes known (Johnston, 1995).

Johnston and others (1994) characterized the seismic potential of unrifted vs. rifted crust in eastern North America using a seismic activity rate normalized to a given unit area of crust. They concluded that the normalized rates for rifted crust in eastern North America are nearly

twice those of unrifted crust. Worldwide, normalized rates for rifted crust exceed those for unrifted crust by a factor of 4 at $M \geq 5$, increasing to a factor of 8 at $M > 6$ (Johnston and others, 1994). This implies that the western boundary of the IRM, and in fact all of the intact Iapetan rifted crust, is very significant in terms of seismic hazard. Therefore, we assign a high weight (0.9) to the model that says the western boundary of the IRM is significant, and we give low weight (0.1) to the model that says it is not significant. In the latter model, the Grenville Front defines the western boundary of the Grenville Domain (GD) source zone, which extends eastward to the southeastern boundary of the IRM source zone (Plate 10).

In southeastern Canada, the Iapetan rifted margin is associated with two failed arms, or aulacogens, that formed transverse to the faulted edge of the ancient continental margin. These aulacogens are oriented at a high angle to the Iapetan margin and extend into the unrifted (non-extended) craton of Grenville crust. They comprise the Saguenay and Ottawa grabens, which are defined by zones of approximately east-west-trending normal faults extending into the Canadian Shield. The normal faults comprising these aulacogens have not undergone much total extension, however, their dimensions and earthquake depths are likely similar to those of the Iapetan margin (Adams and others, 1995). These zones are described in the following sections. In computing the seismicity rates in these zones and the IRM we consider two cases, one in which the aulacogens are separate zones (Plate 1a, 2a, and 3a) and one in which they are part of the IRM (Plate 1b, 2b, and 3b).

3.1.2 Ottawa Graben (OG) Source Zone

The Ottawa graben extends for approximately 700 km into the Canadian Shield (Kumarapeli and Saull, 1966; Kumarapeli, 1985, 1993) from the Sutton Mountains salient of the central Appalachian orogen (Rankin, 1976) (Plate 1). The graben is inferred to extend eastward beneath the Appalachian thrust sheets for approximately 30 km (Kumarapeli, 1993), based on observations that the early Cretaceous Monteregian intrusions (Foland and others, 1986) are emplaced along the graben faults. Alkalic intrusions within the Ottawa graben yield an early Cambrian (~565 Ma) age (Kumarapeli, 1985). A tholeiitic diabase dike swarm associated with the graben is dated at 590 Ma, implying that the initiation of rifting was a late Proterozoic event (Kumarapeli, 1985). The faults defining the Ottawa graben generally strike west-

northwest and offset Silurian strata (Forsyth, 1981). The early Cretaceous Monterigian intrusions at the east end of the Ottawa graben provide evidence for Mesozoic reactivation of the Ottawa rift faults (Kumarapeli and St. Seymour, 1992). In addition, a late Jurassic kimberlite dike occurs at the western end of the rift (Poole and others, 1970). Adams and others (1995) consider the 1935 Timiskaming (M 6.2) event, which occurred at a depth of 10 km, the paradigm earthquake for the Ottawa graben.

Results of the 1982 COCRUST long range seismic refraction experiment show a sharp step-like displacement of the Moho beneath the south shoulder of the graben, confirming the deep-seated nature of its faults and penetration of mantle melts into the crust (Mereu and others, 1986) (Figure 3-3). Furthermore, the COCRUST surveys show a poorly defined Moho at unusually shallow depths beneath the graben (Mereu and others, 1986). These observations suggest the presence of a rift cushion, which would indicate extension at hypocentral depths and throughout the brittle upper crust (Wheeler, 1996c).

3.1.3 Saguenay Graben (SG) Source Zone

The Saguenay graben also represents a failed arm of the Iapetan passive margin (Kumarapeli, 1985), and extends into the Canadian Shield for approximately 300 km from the St. Lawrence Valley (Plate 1). The graben margins are defined by approximately east-west-striking faults; their extent and geometry at depth are unknown (DuBerger and others, 1991). Syn-rift carbonatite complexes dated at 565 Ma imply an early Cambrian age of the graben (Kumarapeli, 1985). Some of the graben bounding faults offset Ordovician limestones by as much as 500 m, and are marked by prominent topographic scarps having 100-300 m of relief (DuBerger and others, 1991). Unlike the Ottawa aulacogen, there is no evidence for Mesozoic reactivation in the Saguenay graben. The paradigm earthquake for the Saguenay graben is the 1988 Saguenay (M 5.9) event, which occurred at a depth of 27 km (Adams and others, 1995).

3.1.4 St. Lawrence Rift Extension (SLE)

As discussed in Section 2.1, various authors have postulated a southwesterly midcontninent extension (i.e., aulacogen) of the St. Lawrence rift system, which is now recognized as a remnant of the Iapetan rifted margin (Wheeler, 1996a). For example, as shown in Figure 2-5,

Adams and Basham (1989) postulated such an extension to the region of the New Madrid rift in the central U.S. Adams and Basham (1991) have since retracted their speculative midcontinent extension of the rift, citing a lack of evidence. To allow for this hypothesis, we consider possible local evidence for a southwesterly extension of the rift system in the Lake Ontario region.

Thomas and others (1993) identified a series of east-northeast-trending linear features in the Rochester Basin of southeastern Lake Ontario from echograms and high-resolution boomer seismic reflection profiles (Figure 3-4). These features were interpreted as faults by Thomas and others (1993), and described as steeply-dipping normal faults, with apparent vertical separations of 10-15 m in the lake bottom glaciolacustrine clays and modern muds. These structures, designated by Thomas and others (1993) as the South Ontario Structural Zone, lie just north of, and parallel to, the southern margin of the postulated southwest extension of the St. Lawrence rift shown by Adams and Basham (1989). Thomas and others (1993) thus interpreted the South Ontario Structural Zone as evidence of a southwest continuation of the St. Lawrence rift into Lake Ontario. They also suggested that the Hamilton-Presqu'ile fault, which lies to the north of, and parallel to, the South Ontario Structural Zone, may represent the northern boundary of the St. Lawrence rift arm extension (Figure 3-1).

Subsequent investigation of part of the South Ontario Structural Zone by McQuest Marine (1995) identified anomalous linear features from airgun seismic reflection profiles plotted at a high (~60x) vertical exaggeration. McQuest Marine (1995) interpreted these anomalous features to be faults, with predominantly down-on-the-north displacements, and apparent throws up to 37.5 m offsetting both bedrock and the lake bottom sediments (Figures 3-5 and 3-6). The downdip extent of these features is unknown. Following the interpretation of Thomas and others (1993), McQuest Marine (1995) suggested these features represent the southern margin of a graben. They further suggested that the northern boundary of the rift lies to the north of the Hamilton-Presqu'ile fault, along the north shore of Lake Ontario (Figure 3-1).

Alternative interpretations of the features comprising the South Ontario Structural Zone have been suggested by Kumarapeli (personal communication cited in Thomas and others, 1993),

Hutchinson and others (1993) and Lewis (1995). Kumarapeli (personal communication cited in Thomas and others, 1993) suggested that the South Ontario Structural Zone features may have formed by collapse due to dissolution of the underlying carbonate bedrock. Hutchinson and others (1993) noted the occurrence of glacial till beneath the lake bottom, and interpreted the linear ridges in the South Ontario Structural Zone to be due to the occurrence of buried drumlins. Lewis (1995) reported that recent high resolution seismic profiles collected in the South Ontario Structural Zone indicate that much of the observed bathymetric relief (10-20 m) is caused by glacial till deposits resting on bedrock. Based on this evidence, Lewis (1995) also interpreted the zone to be a drumlin field. A swath bathymetry map of the area investigated by Lewis (1995) strongly resembles a drumlin field (Figure 3-7).

Wallach (1995b) reported that local outcrop-scale faults extend along the upper St. Lawrence River, from Cornwall to Prince Edward County, and suggested a southwest extending branch of the St. Lawrence rift on the basis of these observations. Wallach (1995b) noted that these locally observed faults appear to coincide with the Hamilton-Presqu'ile fault zone, which extends from Prince Edward County to the western end of Lake Ontario (Ontario Geological Survey, 1991; McFall, 1993), along the postulated northern margin of the rift extension. The amount of displacement on these locally observed faults was not reported by Wallach (1995b).

Grier (1995) reported that brittle faulting occurs on a series of north-northeast and northeast-trending faults along the upper St. Lawrence Valley, from Kingston to Cornwall. The faults in this region generally exhibit offsets of up to 10's of meters, and show variable slip directions (Grier, 1995). However, the amount and timing of brittle deformation on these faults, and their relationship with the much larger faults along the St. Lawrence Valley north of Cornwall is unknown (Grier, 1995).

Gauthier and Benn (1996) identified two prominent northeast-striking faults on the Frontenac arch in northern New York State. These faults are associated with brittle deformation, and are interpreted to be reactivated Grenville ductile structures (Gauthier and Benn, 1996). The recognition of these structures led Gauthier and Benn (1996) to suggest that other northeast-southwest-trending lineaments along the upper St. Lawrence River may correspond to

reactivated basement structures, which may form a possible structural linkage between Lake Ontario and the St. Lawrence rift.

In a recent abstract, Wallach and Thomas (1996) suggested that the Carthage-Colton mylonite zone, a sinuous, ductile structure separating the Frontenac and Central Granulite Blocks within the Grenville province, denotes the southern limit of the St. Lawrence fault zone (i.e., rift). They further suggested that the features in the South Ontario Structural Zone, which lie approximately along the southwest projection of the Carthage-Colton mylonite zone, represent brittle reactivation of this originally ductile Grenville structure.

Deep seismic data collected along and across the Ottawa River (e.g., Mereu and others, 1986) clearly shows crustal-scale extension on the approximately east-west-trending, high-angle faults comprising the Ottawa aulacogen (Figure 3-3). Similarly, seismic profiles collected across the St. Lawrence Valley and the northern Appalachian front indicate step-faulting of the Grenville basement, with displacements ranging from 200-1000 m, extending eastward beneath the Appalachian thrust sheets (St. Julien and others, 1983). The deep seismic data collected in Lakes Ontario and Erie (e.g., Milkereit and others, 1992; Forsyth and others, 1994a,b; Zelt and others, 1994; White and others, 1994) (Figure 3-2) do not show any high-angle extensional structures or extensional deformation on the scale of the Ottawa graben faults or the St. Lawrence Valley faults, providing very strong evidence that a failed rift arm does not extend into the lake. The well-documented, high-angle Iapetan faults that cut Grenville basement along the St. Lawrence, Champlain, Ottawa, and Saguenay river valleys exhibit offsets of the lower Paleozoic rocks on the order of many 100's of meters (e.g., Kumarapeli, 1985; DuBerger, 1991), whereas faulted lower Paleozoic rocks in the northeastern Lake Ontario region (e.g., Grier, 1995; Williams, 1991) generally exhibit maximum offsets on the order of several 10's of meters. Furthermore, unlike the Ottawa, Saguenay, Champlain, and St. Lawrence river valleys, the postulated southwesterly extension of the rift system is not delineated by elevated and persistent seismicity. The origin of the lake bottom features interpreted as evidence for the St. Lawrence rift arm extension is equivocal, and their inferred association with the St. Lawrence rift system is ambiguous at best. The fundamentally different interpretations regarding the origins of these features (i.e., Lewis

(1995) vs. McQuest Marine (1995)) indicates that additional investigations are needed to resolve their true geological nature. Based on the foregoing information, we consider a southwesterly midcontinent extension of the St. Lawrence rift system to be unlikely, and assign a low probability (0.1) to its existence.

Assuming that a southwest-trending midcontinent rift arm does exist, we consider two alternative northern boundaries: the Hamilton-Presqu'ile fault, as suggested by Thomas and others (1993) (Plate 4a, 5a, and 6a); and the northern boundary suggested by McQuest Marine (1995) (Plate 7a, 8a, and 9a). We assign a slightly higher probability (0.6) to the Hamilton-Presqu'ile fault boundary, because it is a mapped structure and traceable for approximately 200 km. The northern boundary proposed by McQuest Marine (1995) is assigned a slightly lower probability of 0.4, because it does not coincide with any mapped features. For both alternatives, we place the southern boundary of the St. Lawrence rift arm extension along the southern margin of the South Ontario Structural Zone, as suggested by both Thomas and others (1993) and McQuest Marine (1995). As was the case for the Ottawa and Saguenay grabens, we consider the extension of the St. Lawrence rift arm to be either a separate source zone (Plate 4a through 9a) or part of the IRM (Plates 4b through 9b).

3.1.5 Extended Continental Crust (ECC) Source Zone

The Extended Continental Crust source zone (ECC) consists of continental crust that most recently experienced significant extension during Mesozoic rifting associated with the separation of the North American and African plates (Plate 1a). This zone includes exposed rift basins along the Atlantic seaboard that are situated landward of the hinge zone of the continental margin. This region experienced considerably less crustal thinning than did the region seaward of the hinge zone that includes the deeper marginal sedimentary basins (Klitgord and others, 1988). The boundaries of this zone generally coincide with the boundaries of the western part of the Eastern Seaboard Domain as defined by Johnston and others (1994) (Figure 2-2). This domain includes Paleozoic basement of part of the eastern United States as well as the eastern continental shelf. The western boundary of this domain, which is taken from Klitgord and others (1988) and Keen and Haworth (1985), generally follows the western edge of the Triassic-Jurassic onshore basins or the boundaries of the

structural blocks in which they occur. As shown by Johnston and others (1994), this zone also includes Paleozoic basins in the northern Appalachians that resulted from post-Acadian transtensional deformation.

The eastern boundary of this zone is defined by the basement hinge zone as shown by Klitgord and others (1988) (Figure 2-6). This zone, which separates shallow platforms from deep marginal sedimentary basins, is a series of parallel half-graben structures that deepen seaward. At the hinge zone, basement deepens steeply to the east by a series of down-dropped fault blocks from about 2 to 4 km depth to over 8 km depth (Klitgord and others, 1988). The character of block faulting at the hinge zone varies along the margin and may reflect the influence of older crustal structure on the Mesozoic rifting (Klitgord and others, 1988). Half-graben structures with seaward-dipping border faults are observed at the Georges Bank hinge zone, whereas faulted blocks with landward-dipping faults form the hinge zone in the Baltimore Canyon trough (Klitgord and others, 1988; Trehu and others, 1989b).

Seismicity within the ECC source zone is not randomly distributed, but tends to be clustered with intervening areas exhibiting low seismicity. The largest historical events within this zone are the earthquakes of 1755 (Cape Ann, **M**~6.2), 1884 (Rockaway Beach/New York City, body wave magnitude, m_b = 5.2), and 1737 (New York City, m_b = 5.2). On a more regional scale, the 1886 Charleston earthquake in South Carolina that is estimated to have been moment magnitude **M** 7.5 occurs within a similar extended continental crust domain (Johnston, 1994).

3.1.6 Great Meteor Hotspot (GMH) Source Zone

The Great Meteor Hotspot source zone (GMH) is adopted from the GAT source zone of Adams and others (1995) (Plate 1a). Adams and Basham (1991) have suggested that the band of seismicity north of the Ottawa River is due to crustal fractures that formed as the North American plate rode over a Cretaceous hotspot (Crough, 1981). Crough (1981) suggested that the passage of the hotspot caused local uplift of the shield, resulting in erosion of at least 1 km in Montreal and as much as 6 to 7 km in New England. Adams and Basham (1991) have postulated that differential uplift of the shield may have thermally stressed and fractured the

Precambrian crust during passage of the hotspot. The GMH source zone lies along the predicted northwest-trending track (Crough, 1981) of this hotspot (Plate 1). The southeastern trace of this hotspot does not appear to be associated with elevated seismicity (Adams and Basham, 1991). This source zone is considered to be distinct even if the IRM/SGR boundary is viewed as insignificant (Plate 10).

3.1.7 Northern Appalachians (NAZ) Source Zone

The Northern Appalachians source zone (NAZ) is a non-extended Paleozoic domain that extends from the southeastern boundary of the IRM source zone to the western limit of Mesozoic extensional faulting (i.e., the western boundary of extended continental crust) (Plate 1). This zone encompasses the northern Appalachian Orogen and comprises sedimentary and metamorphic rocks, and rootless plutons that overrode the Iapetan passive margin. Crustal thicknesses in this region are approximately 40 km (Hughes and Luetgert, 1991). An east dipping zone of detachment (Grenvillian ramp), seen as a major lateral velocity change, reaches a depth of about 25 km in the source zone (Hughes and Luetgert, 1991). This structure has not been identified as a source of recorded seismicity. As noted by Adams and others (1995), all earthquakes in this zone with known depths are relatively shallow (less than 10 km), the paradigm example being the Miramichi earthquake sequence of 1982 (Wetmiller and others, 1984).

3.1.8 Southern/Northern Grenville (SGR/NGR) Source Zones

The southern and northern Grenville source zones (SGR/NGR) represent non-extended crust of the Precambrian Central Craton (Wheeler and Johnston, 1992) of North America (Plates 1, 2, and 3). Basement consists of rocks deformed during the Mesoproterozoic Grenville orogen, which is dated at about 1100 Ma. This zone is characterized by sparse and diffuse seismicity. The Grenville front, which is associated with northwest-directed ductile thrusting, forms the western boundary to this source zone.

3.1.9 Central Craton (CC) Source Zone

The Central Craton source zone (CC) comprises non-extended Precambrian crust of the Eastern Granite-Rhyolite (~ 1450 Ma) and Southern provinces (>2500 Ma) (Plate 1). This zone also is characterized by low rates of seismicity.

3.1.10 Lineament/Seismicity Composite Zones

The regional source zones described above and shown on Plates 1 through 10 are based on subdivision of the region into general "terranes" that reflect a common geology and tectonic history. An alternative approach to identifying homogenous blocks of crust was proposed by Atkinson (1990). Sanford and others (1985) divided the region into a number of crustal blocks based on analysis of satellite fracture mapping and subsurface geologic data. Atkinson (1990) used a systematic analysis of the patterns of shallow (i.e., surfical) and deep (i.e.. seismogenic depths) lineaments together with spatial clustering analyses of seismicity to identify those crustal blocks of Sanford and others (1985) that display similar lineament and seismicity patterns in the site region. She then combined the crustal blocks of Sanford and others (1985) with similar characteristics into the seismic source zones shown on Plate 11. (See Atkinson [1990] for a more detailed discussion.) We include the source zones developed by Atkinson (1990) as an alternative to those based on tectonic history (Plates 1 through 10).

3.1.11 Seismicity-Based Source Zones

The third approach to defining regional source zones is to delineate areas with relatively uniform seismicity. Adams and others (1996) divided the region into a number of source zones that define areas of uniform seismicity. These sources are shown on Plate 13. As discussed in Section 4.1, we also consider a seismicity-based model that relies on spatially varying seismicity rates rather than distinct source zone boundaries to describe regional seismicity patterns.

3.2 Local Seismic Sources

Local seismic sources are defined as sources that extend to within about 100 km of the two nuclear power plant sites. Local sources included in this study provide an opportunity to consider formally a number of tectonic hypotheses regarding the seismogenic potential of

various tectonic features. We have attempted to translate the hypotheses into seismic source characteristics that are appropriate for a probabilistic seismic hazard analysis. The key attributes are: the geometry of the source (defined by the tectonic feature and uncertainties in its location), the probability that the feature is seismogenic, the maximum earthquake magnitude that the source is capable of generating, and earthquake recurrence rates. This section of the report defines the local seismic sources being considered and quantifies the probability that each of them is seismogenic. Sections 4 and 5 discuss the manner in which the sources are modeled, and maximum magnitudes and recurrence defined.

Numerous well-defined linear geophysical anomalies that occur beneath the Paleozoic rocks underlying the Lake Ontario region have been postulated as potentially active seismic sources (Figure 3-1). Many of these anomalies are associated with known, deep-seated Grenville structures, whereas others are interpreted to represent deep-seated basement structures. Other anomalous features observed on the bottom of Lake Ontario (e.g., Thomas and others, 1993) have been interpreted to represent a southwesterly midcontinent extension of the St. Lawrence rift system through the lake. Wallach and Mohajer (1990) and Mohajer (1993) have suggested that reactivated Grenville structures in the Lake Ontario region may be localizing seismicity. Stevens (1995), however, has suggested that seismicity in the Lake Ontario region is characterized by intermittent, scattered activity without any evidence for a spatial association with any particular geologic or geophysical feature. Wallach and Mohajer (1990) reported that the intersections of some of these known and inferred structures, and the possible intersection of these features with the postulated southwest extension of the St. Lawrence rift increases the potential for a moderate to large earthquake in the Lake Ontario region.

In addition to the various geophysical lineaments treated as potential seismic sources, there are numerous small faults in the western Lake Ontario region, some of which displace Quaternary surficial deposits. The origin of these faults is equivocal; Adams and others (1993) have interpreted them as having a nonseismogenic origin because of their short dimensions and other geological characteristics, while Mohajer and others (1992) and Hibbert and others (1994) interpret the structures as having a seismotectonic origin. Because of their recency and equivocal nature, these features are treated in this study as potential seismic sources.

3.2.1 Seismogenic Potential

A key uncertainty in the assessment of local features is whether or not individual faults, lineaments, and other interpreted tectonic features are seismogenic. "Seismogenic" in this context is defined as capable of generating moderate-to-large earthquakes ($m_b \geq 5$). Uncertainty in this assessment, which is specific to each local seismic source, is represented by the relative probabilities on the yes and no branches of the logic tree. In the hazard analysis, the "yes" branch means that the fault or lineament will be treated as a source that localizes seismicity; the "no" branch means that the fault or lineament will not localize seismicity above the levels of the regional source zone within which it lies. In all cases, the regional seismic source zones are assumed to be seismogenic with a probability of 1.0. This is because the regional sources include a large region spatially and likely include at least one fault (although unknown) capable of generating a $m_b \geq 5$ earthquake. Because the regional sources exist everywhere in the study region, there is a finite potential for earthquake occurrence everywhere in the study region. The definition of local sources, then, can define the manner in which future seismicity might be localized along particular tectonic features. The probability that any given tectonic feature is seismogenic varies as a function of evidence for its activity, as discussed below.

The approach that we use in this analysis to quantify explicitly the probability of being seismogenic is essentially the same as that used for the EPRI seismic hazard analysis for the eastern United States (EPRI, 1989). The assessment follows three basic steps:

(1) identification of diagnostic criteria for assessing seismogenic potential and evaluating their relative value,

(2) evaluating individual tectonic features relative to the criteria, and

(3) combining the feature-specific evaluation of the criteria with the relative value of the criteria to arrive at the probability that the tectonic feature is seismogenic. (Note that in the EPRI study, this was termed the probability of "activity").

Below, we first define and rate the criteria, then the procedure for evaluating the probability of being seismogenic is discussed. This is followed by the evaluation of the seismic potential of local sources.

Criteria for Assessing Seismogenic Potential

Given the importance of this component of the source characterization to the hazard analysis, it is useful to define explicitly the criteria that are used to assess whether or not a tectonic feature or lineament is seismogenic, and to define the relative value that each criterion has in making the evaluation. "Relative value", in this sense, is an expression of the resolving power that a particular criterion—and its associated data—has in determining whether or not a tectonic feature is seismogenic. A criterion that provides a high resolving power (i.e., provides a strong indication that a feature is seismogenic) is therefore assigned a high relative weight. For example, if a tectonic feature is spatially associated with several $m_b \geq 5$ earthquakes in the historical record, it would have a high potential for being seismogenic (that is, capable of generating $m_b \geq 5$ earthquakes in the future). Some criteria may be less diagnostic and have a relatively low resolving power for distinguishing whether or not a tectonic feature is seismogenic. An example of such a criterion might be evidence that the tectonic feature has undergone multiple episodes of reactivation during its geologic history. Studies of earthquake occurrence within stable continental regions worldwide (Johnston and others, 1994) have suggested that observed moderate-to-large earthquakes commonly have been associated with multiply-reactivated structures. However, these studies also have concluded that there are many more such tectonic features that are *not* associated with observed earthquakes than have been associated—suggesting that this is not a very diagnostic criterion.

The criteria discussed below are judged to be diagnostic types of information that would indicate whether or not a fault is capable of generating $m_b \geq 5$ earthquakes in the future. These criteria are best suited for potential sources with unclear evidence of activity, but would not be used if certain evidence is well-documented. For example, if a fault has been *causally* (not just *spatially)* associated with large-magnitude historical earthquakes or shows unequivocal evidence of repeated late-Quaternary displacements (as would an active fault in a plate-boundary tectonic setting), then that fault would be considered to be seismogenic with a probability of 1.0. If, however, these most diagnostic criteria are not present for a particular tectonic feature, then the there exist uncertainties in the seismogenic potential of that feature.

This uncertainty is expressed by the probability that the feature is seismogenic and will be less than 1.0. The criteria and their relative diagnostic value are explained in detail below.

Spatial Association with $m_b \geq 5$ Seismicity. The first criterion is the spatial association of a tectonic feature (e.g., fault or lineament) with observed moderate-to-large earthquakes, which have occurred historically or during the instrumental period. Note that this is not a *causal* association (which, clearly, would indicate that the feature is seismogenic), but is merely the spatial distribution of observed earthquakes relative to the feature of interest. We define spatial association as being an *alignment of seismicity along the length of a substantial portion of a given feature* (i.e., a fault or lineament). For general application, the spatial association is made in two-dimensions only (map view), because reliable hypocenter data are not generally available (*See* later discussion for how this type of information in utilized in the assessment). Uncertainties associated with this criterion stem from the fact that, in low activity environments, the historical record is short relative to the recurrence intervals of large earthquakes. As a result, the occurrence of only a single historical event or the absence of large observed events is not uncommon. In addition, it is difficult to associate older historical events with a particular feature because of uncertainties in epicenter location, and sometimes even instrumentally-located events are not easily associated with known faults.

If available, other information including focal depth and focal mechanisms are considered in the final assessment of the seismogenic potential of a given feature. This is applied on a case by case basis, because these data generally are not available for most of the features being considered in this analysis. (*See* discussion below on using feature-specific data sets to "update" the assessment of seismogenic potential).

Spatial Association with $m_b < 5$ Seismicity. The second criterion is the spatial association of a tectonic feature with small-magnitude ($2 \geq m_b < 5$) seismicity. Because seismic networks capable of recording microseismicity ($m_b < 2$) generally are not present in most of eastern Canada, we do not include microseismic events in the definition. Again, we define spatial association as an alignment of seismicity along the length of a substantial portion of a given tectonic feature. It is common for seismogenic faults to be associated with small-magnitude

seismicity, however, the association of a fault or lineament with such events does not necessarily indicate that the feature can also generate larger events.

To illustrate the notion of a "spatial association" of seismicity with a tectonic feature, we present in Figure 3-8 a range of assessments for an idealized linear tectonic feature and idealized "observed" seismicity. The feature C displays a clear association with seismicity aligned along its entire length, and is assessed to be associated with seismicity with a probability of 1.0. In the example, both historical and instrumental events are present—suggesting a persistence through time and little chance that the spatial association with older historical events is merely a matter of the coincidence of poor locations. The feature A in Figure 3-8 has a single recorded event associated with it, and although other earthquakes occur in its vicinity, they show no tendency to align themselves along the feature. Thus, feature A in Figure 3-8 is assessed to be associated with seismicity with a probability of only 0.1.

The other example presented in Figure 3-8 further illustrates the definition of spatial association. Seismicity in the vicinity of feature B on Figure 3-8 occurs primarily near one end. This cluster, however, is part of a trend of seismicity that is perpendicular to the feature and extends away from the feature. Seismicity that occurs near the remaining portion of the feature is scattered, and cannot be distinguished from random background activity. This feature is assessed to have a probability of 0.2 of being spatially associated with seismicity. For two or more features that are proximal to each other, or intersect, we assume that the seismicity in their vicinity can be associated with any of the features. That is, a single seismicity cluster that is proximal to, say three features, is considered in the individual assessments of each of the features.

Multiple Episodes of Reactivation. The third criterion is geological evidence for multiple episodes of reactivation. A fault that exhibits evidence for brittle slip during distinctly different geologic time periods/tectonic phases might be an indication that structure is a persistent zone of weakness in the crust. For example, in their global study of the association of M>4.5 earthquakes with various tectonic features, Johnston and others (1994) found that

several of the moderate-to-large earthquakes that have occurred in stable continental regions have been associated with intra-cratonic rifts that have experienced multiple episodes of reactivation. Admittedly, they also found, however, that there are many rifts displaying such evidence of reactivation that have *not* been associated with seismicity. Uncertainties in applying this criterion arise from the lack of, or uncertainties in cross-cutting relationships of brittle structures, and recognizing different types of cogenetic displacement (e.g., normal faults in the hanging wall of a thrust fault). A structure reactivated in the present day northeast-southwest oriented compressional stress field would also show evidence of activation during a previous phase of deformation.

Brittle Slip in Present Stress Regime. The fourth criterion is geologic evidence for brittle slip that is kinematically consistent with the present tectonic stress regime. Clearly, faults and other features associated with ductile deformation reflect a deformation episode that occurred at a tectonic level that is below the seismogenic zone of the crust. We are, therefore, only considering tectonic features associated with brittle deformation. We consider compressional deformation, i.e., strike-slip, reverse, or oblique reverse faulting, to be an indication of slip within the present day regional stress field. Further, we assume that the axis of maximum horizontal compression is in the northeastern quadrant in the study region. Seismogenic faults, by definition, are favorably oriented relative to tectonic stresses, because earthquakes are a manifestation of the release of tectonic stresses. However, in the absence of other evidence, determination of how favorably a fault is oriented involves a knowledge of the three-dimensional geometry of the fault, the orientations of the principle stress directions, and the rheological properties of the rocks (Zoback, 1992).

Crustal Extent. Because of the rupture dimensions associated with large earthquakes, most seismogenic faults extend through the seismogenic crust (i.e., depths of 10-20 km). The fifth criterion expresses the degree to which a knowledge of the downdip crustal extent of a fault provides an indication of whether or not it is seismogenic. Evidence for expression in the deep crust might include large geophysical anomalies and gradients, the identification of a feature in deep seismic data, and geologic/tectonic interpretations of the tectonic role that a tectonic feature represents (e.g., continental rift faults). Crustal-scale extent is not considered an

exclusively sufficient condition for assessing seismogenic potential. That is, seismogenic tectonic features would be expected to extend to seismogenic depths, but the mere fact that a feature extends to these depths does not provide much assurance that it is in fact seismogenic. This conclusion is supported in many geologic environments, such as much of the eastern United States, where abundant large, crustal- scale faults exist but few are believed to be seismogenic. These features are merely the vestiges of previous tectonic deformation episodes.

Additional Criteria

The criteria discussed above generally are applicable to the tectonic features in the site region because of the available data bases. However, there are additional data of potential use to the evaluation of seismogenic potential that may be available locally, perhaps due to special studies. For example, local seismic networks may provide information on the spatial distribution of microseismicity (m_b<2), or geologic studies may identify evidence for geologically-recent faulting or evidence for paleoseismic shaking (e.g., paleoliquefaction evidence). These observations, although not generally applicable to all local sources, are important to certain local sources and must be taken into account in arriving at the probability of the feature being seismogenic. The additional criteria are discussed here and the procedure for "updating" the assessment to include feature-specific additional data is discussed in the following section.

As defined above, the "spatial association with seismicity" criteria are based on a two-dimensional (map view) association between observed seismicity and the tectonic feature. In some cases, additional seismologic data may exist that could affect the assessment. These data include information on the hypocentral distribution (such that the three-dimensional geometry of the feature can be correlated with seismicity and/or the crustal extent of the feature assessed), focal mechanisms that suggest that the orientation of focal planes are consistent (or not) with the feature of interest, stress tensors that may or may not be consistent with the kinematic indicators on the feature of interest, the spatial pattern of microseismicity (m_b<2), and the spatial pattern of aftershock sequences that define the geometry of coseismic rupture planes.

An additional consideration is geologic evidence for Quaternary tectonic displacement. The Quaternary geologic record spans a much longer period than the historical record, and typically records evidence of large earthquakes that rupture the surface. In plate-boundary tectonic environments, this criterion is the primary means of assessing whether or not particular faults are active. Uncertainties in using this criterion in the site region come from uncertainties in dating the age of slip and discerning whether displacements are tectonic or other in origin. Given the common occurrence of geologically-young glacial features and post-glacial pop-up structures in the Lake Ontario region, the identification of young *tectonic* deformation requires considerable care. In evaluating this criterion for a given fault or lineament, we consider both the geologic evidence for recency of deformation associated with that feature, and the likelihood that the evidence represents seismotectonic or other deformation.

Procedure for Calculating Probability of Being Seismogenic

The diagnostic value of the five seismogenic criteria is shown in Table 3-1 in terms of the conditional probability that a feature is seismogenic given a specific set of observations. The table is laid out in the form of a matrix containing the 32 possible combinations of observing the criteria or lack of the criteria when considering a specific tectonic feature. Each box in the matrix represents one possible set of observations for a tectonic feature. The number in the box represents the conditional probability that the feature is seismogenic given the specific set of observations. The table is arranged such that the upper left hand corner is the most favorable case for the feature being seismogenic (i.e., all five criteria are observed), and the lower right hand corner is the least favorable case for the feature being seismogenic (i.e., all of the criteria are observed to be absent). The criteria are arranged in order of decreasing diagnostic ability from left to right and from top to bottom.

The procedure used to assess the conditional probabilities is to first assess the values at the extremes (i.e., the four corners). For example, in the case where all of the criteria are observed to apply with certainty to a given feature (the upper left hand corner box), the probability that these observations indicate that the feature is seismogenic is judged to be very high (0.95). The reason it is not certainty (1.0) stems from the fact that these observations do not

demonstrate unequivocally a *causal* relationship between the feature and earthquakes. In the case where none of the three geologic criteria is observed for a given feature, yet the feature is spatially associated unequivocally with both large- and small-magnitude seismicity (upper right hand corner box), the probability that the feature is seismogenic also is judged to be very high (0.85).

The bottom row of the matrix in Table 3-1 contains the cases where no spatial association of a tectonic feature with small-magnitude, or moderate-to-large earthquakes is observed. The lower right hand corner is the case where none of the geologic criteria are observed to be present. In this instance, it is judged very unlikely (probability <0.001) that the feature is a localizer of seismicity. Again, the probability is not set to certainty (0) because of our imperfect knowledge of the processes controlling the occurrence of earthquakes in eastern North America. In the case where a feature exhibits unequivocally all of the geological criteria but shows no spatial association with earthquakes, the lower left hand box, the probability of the feature being seismogenic is judged only to be 0.1. This probability reflects our judgement that, in the low-strain tectonic environment of eastern North America, the geologic criteria alone are not strong indicators of the potential of a feature to generate future earthquakes. This is indirectly supported by the widespread occurrence of known geologic structures in the region that have not been associated with historical or instrumental seismicity.

The next stage in assessing the conditional probabilities of being seismogenic is to consider the relative importance of adding (or subtracting) the observation of one of the criteria. Considering the top row of Table 3-1 (when both moderate-to-large and small earthquakes are know to be spatially associated with the feature), the diagnostic power of observing that the feature has deep crustal extent (an increase in conditional probability from 0.85 to 0.87) is judged to be less than the diagnostic power of observing brittle slip (an increase in conditional probability from 0.85 to 0.90), and the diagnostic power of observing multiple reactivations is judged to be greater (an increase in conditional probability from 0.85 to 0.92). Deep crustal extent is judged to be the least diagnostic of the three geologic criteria because some moderate sized earthquakes have occurred in stable continental crust at depths shallower than 10 km. An example is the 1986 m_{bLg} 5 Leroy, Ohio earthquake, which occurred at a depth of about

6 km in Precambrian basement rocks (Seeber and Armbruster. 1993). The same relative importance of the three geologic criteria is used to complete the assessments for the top row and to assess the intermediate conditional probabilities along the bottom row.

The left hand and right hand columns illustrate the relative importance of observed spatial association with moderate-to-large and small magnitude seismicity. The assigned conditional probabilities reflect the judgement that a spatial association with moderate-to-large seismicity is more important that a spatial association with small magnitude seismicity. The assigned probabilities also indicate our judgement that spatial association with seismicity is much more diagnostic of seismic potential than the three geologic criteria.

The remaining conditional probabilities within the interior of Table 3-1 were assigned by interpolating between values along the table margins.

The evaluation of the probability that an individual feature is seismogenic is performed by assessing the degree to which each of the five criteria is judged to apply.

For example, the assessments for a hypothetical feature might be:

CRITERION	PROBABILITY
Association $m_b \geq 5$	0.1
Association $m_b < 5$	0.4
Multiple Reactivation	0.3
Brittle Slip	0.6
Crustal Extent	0.7
Probability of being seismogenic	0.29

The five probabilities listed above indicate the degree to which each criteria is judged to apply to the feature. These probabilities provide a basis for assigning the likelihood that the observed criteria match one of the 32 possible conditions in Table 3-1. The calculation procedure is illustrated in Table 3-2. Each of the five criteria has two possible states: being observed to apply (yes) or being observed to not apply (no). The probabilities listed above are

the probabilities that each criteria is observed (yes), as indicated in Table 3-2. One minus these values are the probabilities that the criteria are not observed. Assuming that the probabilities assigned to observing the five criteria are independent, then the joint probability of a particular set of observed criteria is obtained by multiplying the five individual probabilities. For example, the likelihood that all five criteria are observed is $0.1 \cdot 0.4 \cdot 0.3 \cdot 0.6 \cdot 0.7 = 0.0050$ (first row of Table 3-2) and the likelihood that none of the criteria are observed is $(1-0.1) \cdot (1-0.4) \cdot (1-0.3) \cdot (1-0.6) \cdot (1-0.7) = 0.0454$ (last row of Table 3-2). The unconditional assessment of the probability that the feature is seismogenic is made by multiplying the likelihood that a particular set of criteria is observed by the conditional probability that the feature is seismogenic (Table 3-1) and summing the result over all 32 possible sets of observations (last column of Table 3-2). The result for the example feature is 0.29.

By disaggregating the assessment this way into its component parts, the technical basis for the assessment is made clear. It should be noted, however, that although the assessment is made more explicit by dissecting it into its component parts, the assessment of the probability of the feature being seismogenic is still one that involves professional judgment. The technical basis for the assessments is given for each feature in the following section.

As discussed above, in some cases additional data may exist for certain tectonic features (e.g., focal mechanisms, geologic evidence for Quaternary tectonic slip). In those cases, the probability of being seismogenic that is calculated for the feature—based on the five general criteria—is "updated" to account for the additional information. For example, in the illustration given above the tectonic feature is assessed to have a probability of 0.29 of being seismogenic based on the five general criteria. Suppose that, in addition, a local seismic network has shown that microseismicity is aligned along a significant portion of the tectonic feature, and focal mechanisms for several earthquakes in proximity to the feature indicate that one nodal plane is parallel to the feature. This additional information would lend further support to the seismogenic potential of the feature. The additional criteria are used in a formal Bayesian update of the probability that the feature is seismogenic. The probability of 0.29 assessed from the five primary criteria is considered the "prior" assessment, $P'(S)$. The

"posterior" probability, $P''(S|C)$ given the observation of criteria C is given by Bayes' Theorem

$$P''(S|C) = \frac{P(C|S)\cdot P'(S)}{P(C|S)\cdot P'(S) + P(C|\bar{S})\cdot P'(\bar{S})} \quad \textbf{(3-1)}$$

where $P(C|S)$ is the likelihood (probability) of observing criteria C given that the feature is seismogenic, $P(C|\overline{S})$ is the likelihood of observing criteria C given that the feature is not seismogenic and $P'(\overline{S})$ is the prior probability that the feature is not seismogenic [$P'(\overline{S})$=1.0-$P'(S)$]. If the relative likelihood of observing a criteria when a feature is seismogenic compared to when it is not is defined as $RL(C)=P(C|S)/P(C|\overline{S})$, then Equation **(3-1)** can be rewritten as

$$P''(S|C) = \frac{RL(C)\cdot P'(S)}{RL(C)\cdot P'(S) + (1 - P'(S))} \quad \textbf{(3-2)}$$

If one judges that it is twice as likely that the criteria will be observed for a seismogenic feature than for a non-seismogenic feature, then $P''(S|C) = (2\cdot 0.29)/(2\cdot 0.29+(1-0.29)) = 0.45$. It should be noted that the use of Equation **(3-2)** assumes that the probability of observing the additional criteria given the feature is not active is not zero; that is, the additional criteria is not perfectly diagnostic. If it were, then $P(C|\overline{S})$ would equal zero and the posterior probability would be 1.0 by Equation **(3-1)**.

3.2.2 Local Source Characteristics

This section describes the characteristics of the potential seismic sources within a 100 km radius of the Pickering and Darlington sites, and provides the technical basis for the assessment of their seismogenic potential. Refer to Plates 14 and 15 for the discussion of the local sources. Table 3-3 provides a summary of the assessments for the probability of seismogenicity of the local sources.

Toronto-Hamilton Seismic Zone

Thomas and others (1993) identified a magnetic lineament extending for at least 65 km from Burlington to Toronto, offshore of and parallel to the northeast-trending coastline of western Lake Ontario (Figure 2-7). This lineament, known as the Burlington-Toronto magnetic

lineament (Thomas and others, 1993), and a parallel magnetic lineament to the west, together form a 20-km-wide zone termed by Mohajer (1993) as the Toronto-Hamilton seismic zone. Mohajer (1993) reported that more than 80% of the best-located earthquakes in the western Lake Ontario region are confined to the Toronto-Hamilton seismic zone. Mohajer (1995b) proposed that seismic activity within the zone is "due to its favorable orientation with respect to the current principal compressive stress direction," although this characteristic alone cannot account for its seismic nature. Thomas and others (1993) interpreted anomalous lake bottom features (i.e., dark linear patterns and plumose structures) and bedrock pop-ups observed in the vicinity of the Burlington-Toronto magnetic lineament as evidence of neotectonic activity.

Seismogenic Potential. The following table summarizes the assessment of whether or not the Toronto-Hamilton seismic zone is seismogenic. The individual assessments are described below.

CRITERION (W)	PROBABILITY (P)
Association with $m_b \geq 5$ seismicity	0
Association with $2 \leq m_b < 5$ seismicity	0.3
Seismogenic crustal extent	0.9
Slip favorable w/ present stress regime	0.1
Evidence for reactivation	0.4
Probability of being seismogenic	0.16

The only known historical $m_b \geq 5$ earthquakes in the region are the 1929 m_b 5.2 Attica, New York event, and the 1986 m_{bLg} 5 Leroy, Ohio event. The 1857 **M**~5 Lockport, New York event also occurs in the region, however, there is uncertainty in both its epicentral location and magnitude. Seeber and Armbruster (1993) recently relocated the 1857 event southwest of Buffalo, along the south shore of Lake Erie. Previous interpretations placed this event near Lockport, in upper New York State. In addition to its relocated epicenter, Seeber and Armbruster (1993) revised the magnitude of the Lockport event to **M** 4.

The locations of both the Leroy and Attica earthquakes are well constrained, and the error in their epicentral locations is less than 50 km (Seeber and Armbruster, 1993). Both events occur

well over 50 km away from the Toronto-Hamilton seismic zone, and therefore a probability of zero is assigned to the Toronto-Hamilton seismic zone being associated with $m_b \geq 5$ seismicity.

There is a concentration of small-magnitude seismicity in the western Lake Ontario region relative to the surrounding areas. The depth distribution of instrumentally recorded earthquakes in this region indicates that more than half of the events are located at depths of 5 to 20 km (Mohajer, 1993). Mohajer (1993) reported that eight of the best instrumentally located earthquakes in this region are spatially associated with the Burlington-Toronto magnetic lineament, and that at least thirteen earthquakes are spatially associated with the lineament that defines the western boundary of the zone. However, three of the events associated with the Burlington-Toronto magnetic lineament are just as likely to be associated with the Hamilton-Presqu'ile fault, which intersects the zone to the south (Mohajer, 1993). Earthquakes within the Toronto-Hamilton seismic zone form small clusters that lie partially on and adjacent to the mapped traces of the zone-bounding-lineaments. Seismicity within the south-central part of the zone appears to trend north-south, oblique to the northeast trend of the zone. The Southern Ontario Seismic Network (Mereu and others, 1996) has recorded nine events ($m_b>1$) within the southern part of the zone since the network began operation in 1991. A number of these events form a northeast-trending alignment which lies within, and approximately parallel to the southern part of the Toronto-Hamilton seismic zone. Given the concentration of small-magnitude seismicity in this region, and the partial alignment of seismicity along the south-central part of the Toronto-Hamilton seismic zone, we assign low-to-moderate probability (0.3) to the Toronto-Hamilton seismic zone being associated with small-magnitude seismicity. Mohajer (1993) has also recorded small magnitude seismicity in the area.

It is unknown if the Toronto-Hamilton seismic zone is correlative to a structure at depth. Deep seismic data collected in Lake Ontario (e.g., Forsyth and others (1994a) did not cross this feature. Eyles and others (1993) suggested that the zone coincides with the Fishog-Cambridge terrane boundary within the Grenville basement, though, Carter and Easton (1990) interpret this boundary to lie considerably west of Lake Ontario (Figure 3-9). Instrumentally recorded seismicity within the Toronto-Hamilton seismic zone does extend to a depth of about 20 km

(Mohajer, 1993). Given the prominent magnetic signature of the feature and the deep extent of the seismicity, a very high probability (0.9) is assigned the Toronto-Hamilton seismic zone extending to seismogenic depths.

There is no clear geologic evidence for brittle tectonic faulting within the Toronto-Hamilton seismic zone in the present tectonic stress regime. Although west-northwest-trending pop-up structures in the vicinity of the Burlington-Toronto magnetic lineament indicate buckling of strata on the bedrock surface consistent with the contemporary northeast-oriented horizontal compressive stress, these features are not considered to be spatially or structurally associated with the Toronto-Hamilton seismic zone. Lewis (1995) and Lewis and others (1995) have shown that these features are widely distributed on the flanks of the Lake Ontario basin, and are spatially unrelated to the Burlington-Toronto magnetic lineament and the various other geophysical lineaments beneath the lake. Thomas and others (1993) also observed plumose structures and dark linear features on the bottom of western Lake Ontario and interpreted these features as indicators of neotectonic activity. They suggested that the plumose structures represent the surface expression of fracture or fault propagation, and that the dark linear patterns result from structurally-controlled gas injections into the lake bottom sediments (Thomas and others, 1993). While processes of this type could indicate the presence of open joints or fractures in the underlying bedrock, they would not necessarily imply recent brittle tectonic faulting. Lewis and others (1995) have reported that the distribution and character of both the linear patterns and plumose structures do not fully support tectonic origins for the features. Furthermore, Lewis and others (1995) have suggested anthropogenic origins for both sets of features, namely: that the plumose structures represent anchor drag marks, and that the dark linear patterns may represent steamship ash debris trains. Considering the lack of clear evidence for brittle tectonic faulting within the Toronto-Hamilton seismic zone, we assign a very low probability (0.1) to this criterion.

There are no data that can be used to assess if multiple episodes of brittle reactivation occurred in the Toronto-Hamilton seismic zone. Therefore, we assign a moderate probability of 0.4 to reflect the judgment that fewer structures display evidence of multiple reactivations than those that display evidence of no reactivation.

Based on our assessment, we arrive at the probability of 0.16 that the Toronto-Hamilton Seismic Zone is seismogenic (*See* Section 3.2.1 for calculation procedure).

The Niagara-Pickering Linear Zone

The Niagara-Pickering linear zone is a north-northeast-trending feature defined by a prominent linear aeromagnetic anomaly, a less well-defined Bouguer gravity anomaly, and linear physiographic features (Wallach and Mohajer, 1990). The Niagara-Pickering linear zone coincides with the southward projection of the Central Medisedimentary Belt boundary zone (CMBBZ), which is exposed in the Grenville Province and extends southward from the Canadian Shield beneath the overlying unconformable lower Paleozoic cover (Figure 3-1). The depth to the magnetic sources along strike of the Niagara-Pickering linear zone gradually increases from 2 km in the north to 4 km in the south, consistent with the southward thickening of the Paleozoic cover (Roest, 1995). Roest (1995) reported that the magnetic field data indicate the Niagara-Pickering magnetic lineament is not an isolated feature. He described it as a series of more or less en-echelon, curvilinear anomalies within a zone (i.e., the CMBBZ) that widens to the north and south of the northern Lake Ontario shoreline. Roest (1995) also reported that it is not the highest amplitude anomaly over Lake Ontario, and that the associated gravity signature is relatively subtle.

The CMBBZ is a major east-dipping ductile terrane boundary separating rocks of the Central Metasedimentary belt on the east from those of the Central Gneiss belt within the Grenville Orogen (Hanmer, 1988). South of Lake Ontario, the magnetic anomaly associated with the CMBBZ/Niagara-Pickering linear zone is postulated to connect with the Akron magnetic boundary (Wallach, 1990; Mohajer, 1993), a candidate seismogenic feature (Seeber and Armbruster, 1993) that extends beyond the south shore of Lake Erie and is thought to represent the same Grenville terrane boundary (Rankin and others, 1993).

Wallach (1995a) reported that outcrop-scale faults and fractures within Grenvillian and upper middle Ordovician rocks within and adjacent to the Niagara-Pickering linear zone /CMBBZ provide unequivocal evidence for three separate phases of brittle reverse, normal and strike-slip faulting along the Niagara-Pickering linear zone/CMBBZ and a zone at least 30 km-wide

to the east. However, he did not report on the amount or timing of faulting, or the full extent of the area affected. Wallach (1995a) also reported evidence of macro-scale faulting along the Niagara-Pickering linear zone/CMBBZ, based on an apparent right-lateral separation of lower Paleozoic rocks shown on a bedrock geology map of Ontario (Ontario Geological Survey, 1991). Wallach and Mohajer (1990) suggested that the possible spatial association of historical earthquakes with the Niagara-Pickering linear zone, including the 1853 MM intensity V event, and possibly the 1873 MM intensity VI event, suggests that the Niagara-Pickering linear zone may be seismically active.

Seismogenic Potential. There are uncertainties associated with whether or not the Niagara-Pickering linear zone is associated with the Rouge River faults, and whether or not it is associated with the Akron magnetic boundary. (The logic structure for evaluating these possible associations is discussed in Section 4 and is shown in Figure 4-2). Because the possibility of these assessments affects the assessment of seismogenic potential, four assessments are made for the Niagara-Pickering linear zone. The four assessments are as follows:

(1) assumes an association with the Rouge River faults and no association the Akron magnetic boundary;

(2) assumes no association with the Akron magnetic boundary or the Rouge River faults;

(3) assumes an association with the Akron magnetic boundary and no association with the Rouge River faults;

(4) assumes an association with the Rouge River faults and with the Akron magnetic boundary.

Mohajer (1995a) suggested that the west-northwest-trending Rouge River faults result from reactivation of basement structures in the region and identified the CMBBZ as the probable causative structure. The structures comprising the Rouge River faults trend nearly perpendicular to the north-northeast trend of the Niagara Pickering linear zone, nevertheless,

we consider the possibility of an association in our assessment because of the close proximity of these features. Wallach (1995a) suggested that the Rouge River faults may be associated with the northwest-trending Georgian Bay linear zone because their trends are similar, and because the Rouge River faults lie within the approximately 50-km-wide Georgian Bay linear zone. Because of these considerations, we also consider the Rouge River faults to be possibly associated with the Georgian Bay linear zone. However, because of the short dimensions of the Rouge River faults, their normal sense of displacement, observations supporting nonseismogenic origins of the features (e.g., Adams and others, 1993), and the equivocal involvement of bedrock faulting, they likely are not associated with a deep-seated, crustal-scale tectonic fault. Rather, their characteristics suggest that they likely are separate, enigmatic features. Therefore, we assign higher weight (0.7) to the model that says they are a separate source, and we assign lower weight (0.15) to the models that treats them as part of the Niagara-Pickering linear zone or the Georgian Bay linear zone.

There also are uncertainties associated with whether or not the CMBBZ/Niagara-Pickering linear zone is structurally associated with the Akron magnetic boundary (i.e., whether or not the two features represent the same Grenville terrane boundary) (Plate 16). It has been postulated that the two features represent the same Grenville tectonic boundary, based on similarities in their magnetic attributes (Rankin and others, 1993), and the observation that the lineaments appear to curve towards one another (Wallach, 1995a). Wallach (1995a) also noted that the apparent curvature evident in the magnetic field map also is evident on the gravity map of the region. Rankin and others (1992) have interpreted the Akron magnetic boundary as a major Grenville terrane boundary, and they have shown the northern extension of the Akron magnetic boundary as a continuous feature that appears to coincide with the CMBBZ (Figure 3-10). Carter and Easton (1990) and Easton (1992), however, have not interpreted the magnetic features as representing the same structural boundary. Carter and Easton (1990) utilized both magnetic and limited borehole data in southern Ontario to extend the CMBBZ beneath the Paleozoic cover as far south as northeastern Pennsylvania (Figures 3-9 and 2-8). Forsyth and others (1994a) also used structural constraints, along with magnetic data to interpret the southerly continuation of the CMBBZ beneath western Lake Ontario and eastern Lake Erie, and into northeastern Pennsylvania, approximately coincident with the

interpretation of Carter and Easton (1990) (Figure 2-11). Given that the interpretations of Carter and Easton (1990) and Forsyth and others (1994a) are based on both magnetic and geologic data, we assign higher weight (0.6) to the model that says the CMBBZ and the Akron magnetic boundary are not the same tectonic feature, and we assign lower weight (0.4) to the model that says they are the same tectonic feature. In the case where the two features are assumed to be connected (i.e., the same tectonic feature), the geometry of the Akron magnetic boundary is assumed to be consistent with the geometry of the Niagara-Pickering linear zone/CMBBZ as imaged in the deep seismic data (e.g., White and others, 1994).

The following tables summarize the assessments of the seismogenic potential of the Niagara Pickering linear zone. The individual assessments are characterized below.

Assuming an association with the Rouge River faults and no association with the Akron lineament:

CRITERION (W)	PROBABILITY (P)
Association with $m_b \geq 5$ seismicity	0
Association with $2 \leq m_b < 5$ seismicity	0.1
Evidence for reactivation	0.6
Slip favorable w/ present stress regime	0.3
Seismogenic crustal extent	1.0
Probability of being seismogenic	0.09

No association with the Rouge River faults and no association with the Akron lineament:

CRITERION (W)	PROBABILITY (P)
Association with $m_b \geq 5$ seismicity	0
Association with $2 \leq m_b < 5$ seismicity	0.1
Evidence for reactivation	0.5
Slip favorable w/ present stress regime	0.2
Seismogenic crustal extent	1.0
Probability of being seismogenic	0.08

Assuming an association with the Akron magnetic lineament and no association with the Rouge River faults:

CRITERION (W)	PROBABILITY (P)
Association with $m_b \geq 5$ seismicity	0.2
Association with $2 \leq m_b < 5$ seismicity	0.25
Evidence for reactivation	0.5
Slip favorable w/ present stress regime	0.2
Seismogenic crustal extent	1.0
Probability of being seismogenic	0.30

Assuming an association with the Akron magnetic lineament and the Rouge River Faults:

CRITERION (W)	PROBABILITY (P)
Association with $m_b \geq 5$ seismicity	0.2
Association with $2 \leq m_b < 5$ seismicity	0.25
Evidence for reactivation	0.6
Slip favorable w/ present stress regime	0.3
Seismogenic crustal extent	1.0
Probability of being seismogenic	0.31

The 1929 m_b 5.2 Attica, New York event, and the 1986 m_{N_g} 5 Leroy, Ohio event are the only known m_b 5 earthquakes in the region that could have occurred on the Niagara-Pickering linear zone. The epicentral location of the Attica event is well-constrained, and lies 80 km to the southeast of the Niagara-Pickering linear zone. Given the error in the epicentral location of the Attica event (<50 km), it cannot be associated with the Niagara-Pickering linear zone. The Leroy event is located well over 100 km from the Niagara-Pickering linear zone. When the Niagara-Pickering linear zone and the Akron magnetic boundary are not connected, a probability of zero is assigned to the Niagara-Pickering linear zone being associated with $M_b 5$ seismicity. The epicentral location of the m_{N_g} 5 Leroy event is well-constrained, and appears to be spatially associated with the Akron magnetic boundary (Seeber and Armbruster, 1991, 1991). Two additional $m_b 5$ events (one historical and the other recorded) occur proximal to the Akron magnetic boundary in southern Ohio (Plate 12). Therefore, when a connection

between the Niagara-Pickering linear zone and the Akron magnetic boundary is assumed, a low probability (0.2) is assigned to the Niagara-Pickering linear zone being associated with $m_n \geq 5$ seismicity.

A cluster of historical and instrumentally recorded small-magnitude seismicity occurs in the Niagara Falls area, between Lakes Ontario and Erie. Very few of the events within this cluster are spatially coincident with the Niagara-Pickering linear zone. Along the northern portion of the Niagara-Pickering linear zone, seismicity is much more diffuse, and even fewer instrumentally recorded small-magnitude events occur in the vicinity of the mapped trace of the Niagara-Pickering linear zone/CMBBZ. Deep seismic data (e.g. Forsyth and others, 1994a) indicate that the CMBBZ is a gently east-dipping structure. Considering this geometry and the depth distribution of earthquakes in the western Lake Ontario region, events located near the surface trace that occur deeper than about 5 km are likely not associated with this feature. East of the mapped trace (i.e. downdip) seismicity is diffuse and cannot be distinguished from random background activity. For the case when there is no connection between the Niagara-Pickering linear zone and the Akron magnetic boundary, we assign a very low probability (0.1) to the Niagara-Pickering linear zone being associated with small-magnitude seismicity. This assessment reflects the observation that very few events are spatially associated with the Niagara-Pickering linear zone, and there is no associated linear alignment of small-magnitude seismicity along it.

A relatively dense cluster of small-magnitude seismicity occurs in the vicinity of the Akron magnetic boundary. A number of these events, both historical and recorded, are aligned along the trend of the feature. However, since the associated seismicity forms an alignment along only a small portion of the entire feature, we assign a low probability (0.25) to the Niagara-Pickering linear zone being associated with small-magnitude seismicity.

Deep seismic reflection profiles collected across the CMBBZ in Lakes Erie and Ontario show that this gently east-dipping feature extends to the lower crust. Further, the CMBBZ is a profound feature in the magnetic data, suggesting a deep crustal structure. Therefore, a very

high probability (1.0) is assigned to the Niagara-Pickering linear zone/CMBBZ extending to seismogenic depths.

Evidence for brittle slip on the Niagara-Pickering linear zone in the contemporary tectonic stress regime consists of the meso-scale brittle reverse and strike-slip faults along the CMBBZ/Niagara-Pickering linear zone reported by Wallach (1995a), and an apparent right lateral separation of the upper middle Ordovician rocks at Lake Scugog shown on the bedrock geology map of southern Ontario (Ontario Geological Survey, 1991). The apparent right lateral separation at Lake Scugog is not compelling evidence of brittle faulting. Much of the regional bedrock distribution was determined from waterwell records, and there is very limited data coverage in this area (Eyles and others, 1993). Further, there is no corresponding right lateral offset in the middle Ordovician outcrop belt to the north (Ontario Geological Survey, 1991). Hence, the mapped bedrock distribution, if true, could be explained simply by glacial erosion.

Sanford (1993) examined the middle Ordovician rocks along the Precambrian/Paleozoic contact at the western boundary of the CMBBZ and found no evidence of brittle faulting at this location. Sanford (1993) did document normal faulting at several locations to the east of this area. The faults were predominantly northwest-striking and generally exhibited vertical offsets of less than one meter. Sanford (1993) noted minor northwest-striking strike-slip faults at one location, and a normal fault with a vertical displacement of several 10's of meters at another location. Sanford (1995) and McQuest Marine (1995) also interpreted possible faults beneath Lake Ontario in the vicinity of the intersection between the Niagara-Pickering linear zone and the Hamilton-Presqu'ile fault. However, these interpreted faults are oriented approximately parallel to the Hamilton-Presqu'ile fault, and therefore, we consider an association of these features with the Niagara-Pickering linear zone to be unlikely. Lewis and others (1995) documented numerous small displacement (0.5 m) normal faults in this area, however, these structures were limited to a package of stratified sediment and did not penetrate bedrock. Lewis and others (1995) suggested the structures may have formed due to differential settling of the underlying sediments. Lewis and others (1995) also documented numerous sediment depressions in the vicinity of the Niagara-Pickering linear zone, the

margins of which resemble monoclines. These features were attributed to sediment drape into a depression, rather than displacement of sediment after deposition, based on continuous and undisturbed reflection data (Lewis and others, 1995). The depressions could not be correlated to deeper sedimentological or bedrock features, which strongly suggests a nontectonic origin (Lewis and others, 1995). Lewis and others (1995) noted that these features widely distributed in the basin, and are spatially unrelated to the various underlying geophysical lineaments. Based on the available evidence, which includes the local reverse and strike-slip faults in the middle Ordovician rocks reported by Wallach (1995a), and the minor strike-slip faults in the vicinity of the Niagara-Pickering linear zone documented by Sanford (1993), we assign a low probability (0.2) to the Niagara-Pickering linear zone having evidence of brittle slip in the present tectonic stress regime.

Mohajer (1995a) and Mohajer and others (1995) have suggested that the Quaternary Rouge River faults represent reactivation of the CMBBZ. However, the normal faults identified by Mohajer and others (1992) and Hibbert and others (1994) trend west-northwest, which is orthogonal to the north-northeast-trending structures comprising CMBBZ. Furthermore, normal faulting is not consistent with the present day compressional tectonic stress field. Given their orientations and normal sense of displacement, a low-to-moderate probability (0.3) is assigned to the Niagara-Pickering linear zone having evidence for recent brittle slip if an association with the Rouge River faults is assumed.

Wallach (1990, 1995a) reported observations of outcrop scale structures in both Precambrian and Paleozoic rocks, and suggested that the Niagara-Pickering linear zone/CMBBZ has undergone several episodes of brittle reactivation. Deep seismic reflection data collected in eastern Lake Erie indicate the CMBBZ was reactivated once in extension (Milkereit and others, 1992) (Figure 3-2). Eyles and others (1993) presented evidence suggesting a possible relationship between the CMBBZ and a relict drainage system cut across the lower Paleozoic cover in the western Lake Ontario region. Eyles and others (1993) suggested that the position of a northeast-trending channel on the Niagara Peninsula, along with another northeast-trending channel north of Lake Ontario, indicates the upward propagation of fractures associated with reactivation of the CMBBZ. Wallach and Mohajer (1990) also noted that

northeast-trending linear lakes and swamps parallel the CMBBZ/Niagara-Pickering linear zone.

Sanford (1993) and Rutty and Cruden (1993) documented north-northeast, as well northeast and northwest-trending, surface lineaments along the CMBBZ/Niagara-Pickering linear zone in both the lower Paleozoic rocks and the exposed Grenville rocks. The origins of these north-northeast-trending surface lineaments is questionable, and it is not known whether they are geographically limited to the CMBBZ/Niagara-Pickering linear zone (Sanford, 1993; Rutty and Cruden, 1993). North-northeast-trending joints and fractures have been documented in numerous other areas around Lake Ontario (e.g., Rogojina and others, 1993; McFall and Allam, 1991; Mohajer and others, 1992). Rutty and Cruden (1993) also reported that northwest-trending pop-ups are the only neotectonic structures present within the Paleozoic rocks immediately south of the exposed CMBBZ, and that their orientations suggest they are genetically unrelated to recent reactivation of the CMBBZ.

From the deep seismic data it is clear that the CMBBZ has undergone at least one episode of reactivation. The spatial association of the relic Paleozoic drainages and the CMBBZ also suggests possible reactivation, though it is not as compelling as the seismic data. It is not clear whether the local brittle faulting in the Grenvillian and lower Paleozoic rocks reported by Wallach (1995a) and Sanford (1993) is associated with reactivation of the CMBBZ or other processes, such as isostatic depression and rebound. Sanford (1993) suggested that periodic uplift of the regional basement-cored arches during the various Phanerozoic compressional and extensional tectonic regimes could be responsible for the small-scale deformation observed in the lower Paleozoic rocks throughout the southern Ontario region. Small-scale faulting occurs throughout the Ordovician outcrop belt north of Lake Ontario, and these structures have not been systematically analyzed to determine if they are anomalous in the vicinity of the Niagara-Pickering linear zone. Considering the foregoing information, we assign a moderate probability (0.5) to the Niagara-Pickering linear zone/CMBBZ having evidence of multiple episodes of reactivation. As mentioned previously, Mohajer (1995a) and Mohajer and others (1995) have suggested that the faults identified along the Rouge River represent brittle reactivation of the CMBBZ. However, given their west-northwest

orientations, which is orthogonal to the CMBBZ, and their normal sense of displacement, we do not consider them to be strong evidence of reactivation of this north-northeast-trending feature. Therefore, if an association with the Rouge River faults is assumed, the probability of the NPLZ having evidence of multiple episodes of reactivations is increased only slightly from 0.5 to 0.6.

Following the logic given in Section 4 of this report and shown in Figure 4-2, when considering the possible association of the Rouge River faults with the Niagara-Pickering linear zone, we arrive at a probability of 0.09 that the Niagara Pickering linear zone is seismogenic. When we do not consider an association between the Rouge River faults and the Niagara Pickering linear zone, we arrive at a probability of 0.08 that the Niagara Pickering linear zone is seismogenic. The combined probability that the Niagara Pickering linear zone is associated with the Rouge River faults, and the Niagara Pickering linear zone is seismogenic is (0.15 x 0.09 = 0.014). Likewise, the combined probability that the Rouge River faults are *not* associated with the Niagara Pickering linear zone, and the Niagara Pickering linear zone is seismogenic is (0.85 x 0.08 = 0.068).

If a connection between the Niagara Pickering linear zone, the Akron magnetic boundary, and the Rouge River faults is assumed, we arrive at a probability of 0.31 that the Niagara Pickering linear zone is seismogenic. If a connection between the Niagara Pickering linear zone and the Akron magnetic boundary, without association with the Rouge River faults is assumed, we arrive at a probability of 0.30 that the Niagara Pickering linear zone is seismogenic. The combined probability that the Niagara-Pickering linear zone is associated with the Akron magnetic boundary and the Rouge River faults, and the Niagara-Pickering linear zone is seismogenic is (0.4 x 0.15 x 0.31 = 0.119). Likewise, the combined probability that the Niagara-Pickering linear zone is associated with the Akron magnetic boundary and *not* with the Rouge River faults, and the Niagara-Pickering linear zone is seismogenic is (0.4 x 0.85 x 0.30 = 0.102).

Additional Information. Additional information exists that is pertinent to the assessment of the seismogenic potential of the Niagara-Pickering linear zone. Aftershock hypocenters of the

1986 m_{bLg} 5 Leroy, Ohio event suggest a northeast-striking, vertical rupture about 1 km across, and place the event at a depth of about 6 km within the Grenville basement (Seeber and Armbruster, 1993, 1995). This orientation is parallel to the trend of the Akron magnetic boundary and lends support to the possibility that the earthquake was associated with the feature. Fault plane solutions suggest a right-lateral motion, which is consistent with the northeast-southwest orientation of horizontal compression in the regional stress field (Seeber and Armbruster, 1993, 1995).

If the Niagara-Pickering linear zone/CMBBZ and the Akron magnetic boundary are assumed to be a single source, the geometry of the feature is assumed to be consistent with the geometry of the CMBBZ as determined from deep seismic data (e.g., White and others, 1994; Forsyth and others, 1994a). These data indicate that the north-northeast-trending CMBBZ comprises gently east-dipping structures (White and others, 1994). Focal mechanisms for the Leroy event yield a northeast-striking, vertical plane, which is not consistent with the geometry of the CMBBZ. In our judgement, the relative likelihood of observing an inconsistent focal mechanism for a given seismogenic feature compared to a given nonseismogenic feature is estimated to be 1.25:1.

In addition to this information, the Rouge River faults are considered as possible evidence of Quaternary tectonic faulting. The relative likelihood of observing faulted Quaternary deposits in southern Ontario along a nonseismogenic feature compared to a seismogenic feature is judged to be 1. That is, we consider the probability of observing offset Quaternary deposits along a given seismogenic feature and a given nonseismogenic feature to be equally likely. This judgment reflects the observations that (1) glacially deformed Quaternary sediments are recognized locally throughout the region (e.g., van der Wateren, 1987; Broster, 1991), (2) there are few known seismogenic structures in southeastern Canada, and (3) there is only one known historical surface rupture in eastern Canada (e.g., Adams, 1996). Because the relative likelihood is 1:1, the observation of Quaternary faulting provides no additional information to update the probability that the feature is seismogenic.

Given the additional information and using Equation (3-2), the probability that the Niagara-Pickering linear zone is seismogenic is updated from 0.31 to 0.36 when an association with the

Akron magnetic boundary and the Rouge River faults is assumed. Similarly, when an association with the Akron magnetic boundary and no association with the Rouge River faults is assumed, the probability that the Niagara-Pickering linear zone is seismogenic is updated from 0.30 to 0.35.

The Wilson-Port Hope Magnetic Lineament

The northeast-trending Wilson-Port Hope magnetic lineament extends for approximately 80 km across central Lake Ontario, from Port Hope, Ontario in the north, to Wilson, New York in the south (Mohajer, 1993; McQuest Marine, 1995) (Figure 3-1). It lies between, and subparallel to, the Niagara-Pickering linear zone and the Clarendon-Linden fault system, along the trend of a system of prominent subsurface fractures in the Lake Erie-Niagara Falls area identified by Sanford and others (1985) (McQuest Marine, 1995). Mohajer (1993) showed the Wilson-Port Hope lineament as comprising three discontinuous, slightly curvilinear segments. McQuest Marine (1995) suggested this feature may correspond to a Precambrian basement structure. Although recent seismic reflection data did not reveal any faults along this lineament, McQuest Marine (1995) concluded that evidence from other studies (e.g. Sanford and others, 1985) suggests that the Wilson-Port Hope lineament may be a currently active fault.

Seismogenic Potential. The following table summarizes the assessment of the seismogenic potential of the Wilson-Port Hope lineament. The individual assessments are described below.

CRITERION (W)	PROBABILITY (P)
Association with $m_b \geq 5$ seismicity	0
Association with $2 \leq m_b < 5$ seismicity	0.1
Evidence for reactivation	0.4
Slip favorable w/ present stress regime	0.1
Seismogenic crustal extent	0.2
Probability of being seismogenic	0.06

Both of the known $m_b \geq 5$ earthquakes in the region (1929 m_b 5.2 Attica, New York, 1986 m_{bLg} 5 Leroy, Ohio) are located more than 50 km away from the Wilson-Port Hope lineament. Similarly, the relocated epicenter of the 1857 event lies approximately 75 km southwest of the lineament. Considering that the errors in the epicentral locations of these events are less then 50 km, a probability of zero is assigned to the Wilson-Port Hope lineament being associated with $m_b \geq 5$ seismicity.

There is no obvious spatial association of the Wilson-Port Hope lineament with small-magnitude seismicity. A few events occur nearthe northern end of the lineament, however, and could be projected onto it. Several events also occur near the intersection of this lineament with the Georgian Bay linear zone, and these events could be associated with either feature. Based on the scarcity of events that occur near the Wilson-Port Hope lineament and the lack of an alignment of events along this feature, we assign a very low probability (0.1) to the Wilson-Port Hope lineament being associated with small-magnitude seismicity.

It is not known if this magnetic feature corresponds to an actual structure in the Grenville basement, or whether it simply reflects variablebasement lithologies. It does not appear to be defined by truncated, offset potential field anomalies, rather, it appears to be defined by a linear, narrow positive anomaly. Although its tend is subparallel to other known Grenville structures in the region, a structure corresponding to this anomaly is not readily apparent in the data of Forsyth and others (1994a). Therefore, a low probability (0.2) is assigned to the Wilson-Port Hope lineament extending to seismogenic depths.

Seismic reflection data collected across the Wilson-Port Hope lineament in southern Lake Ontario did not reveal the presence of any faultsin the lake bottom (McQuest Marine, 1995). Sidescan sonar lines collected across this lineament revealed dark linear features, some interpreted as anchor scours, and segmented dar marks interpreted to be possible degassing marks along fractures (McQuest Marine, 1995). The distribution of these features does not suggest a spatial association with the lineament. McQuest Marine (1995) also reported that a system of open fractures at least 40 km long ocurs roughly perpendicular to the southern end of the Wilson-Port Hope lineament, based on sediment-water-chemical relationships that have

been identified by Drimmie (personal communication cited in McQuest Marine, 1995). However, the relationship between the inferred open fracture system and the Wilson-Port Hope lineament is ambiguous, and no explanation is offered by McQuest Marine (1995). Given the lack of clear evidence for brittle faulting on the Wilson-Port Hope lineament, but considering the possibility that the lake bottom features identified by McQuest Marine (1995) represent slip at depth and are associated with the Wilson-Port Hope lineament, we assign a very low probability (0.1) to the Wilson-Port Hope lineament having evidence for brittle slip consistent with the present day regional stress regime.

There is no evidence that the Wilson-Port Hope lineament has undergone multiple episodes of reactivation. Therefore, we assign a moderate probability of 0.4 to reflect the judgment that fewer structures display evidence of multiple reactivations than those that display evidence of no reactivation.

Based on the above assessments, the probability that the Wilson-Port Hope lineament is seismogenic is 0.06.

Clarendon-Linden Fault System

The Clarendon-Linden fault system (Van Tyne, 1975; Fakundiny and others, 1978) of upper New York State comprises a buried zone of predominantly north-northeast-striking faults that can be traced for approximately 150 km from just north of the Pennsylvania border (Jacobi and Fountain, 1993) to the north shore of Lake Ontario (Hutchinson and others, 1979) (Figure 3-1). Closely-spaced, small-offset step faults characterize the Clarendon-Linden fault system in outcrops of Devonian rocks (Jacobi and Fountain, 1993, 1996). Jacobi and Fountain (1996) have suggested that monoclines observed on old seismic data could actually be a series of small step faults that were previously not imaged because of low resolution seismic data. Recent seismic reflection data show Clarendon-Linden faults extending very close to the surface (where all coherent reflectors are lost) (Jacobi and Fountain, 1996). The main strand of the Clarendon-Linden fault system is highly segmented, steeply east-dipping in the north and west-dipping in the south, and displays a maximum vertical displacement of about 80 m (Jacobi and Fountain, 1993). Cumulative offset across the Clarendon-Linden fault system is

as high as 130-200 m, based on well log, outcrop stratigraphic and seismic data (Jacobi and others, 1996). Forsyth and others (1994b) have suggested that the Clarendon-Linden fault system is part of a wider zone of small Paleozoic faults that lie above the crestal area of a northeast-trending Precambrian bedrock high, termed the Iroquonian high.

The Clarendon-Linden fault system is inferred to extend northeastward beneath Lake Ontario, coincident with a bathymetric lineament, known as the Scotch Bonnet Rise (Hutchinson and others, 1979). The Scotch Bonnet Rise is a west-facing bedrock ridge that exhibits approximately 20 m of relief and may be related to faulting (Anderson and Lewis, 1975). Hutchinson and others (1979) collected seismic data across this feature and, within the resolution of their data (~2-3 m), found no evidence for post-glacial Holocene faulting across the Scotch Bonnet Rise. Both the Clarendon-Linden fault system and the Scotch Bonnet Rise coincide with the east flank of a magnetic anomaly, and the west edge of a series of Bouguer gravity anomalies, both of which can be traced from west-central New York to the north shore of Lake Ontario (Hutchinson and others, 1979). Deep seismic reflection data suggest that the Clarendon-Linden fault system is coincident with structures associated with the Elzevir-Frontenac terrane boundary zone of the Grenville province (Forsyth and others, 1994b; Easton and others, 1995). Seismic reflection profiles clearly show the gently east-dipping ductile thrusts of the Elzivir-Frontenac boundary zone extending up to the Precambrian/Paleozoic contact in the region of the Clarendon-Linden fault system (Forsyth and others, 1994; Jacobi and Fountain, 1996). However, the Precambrian structures do not lead directly to individual Clarendon-Linden faults (Jacobi and Fountain, 1996). The Salmon River fault, which is exposed approximately 30 km north of Lake Ontario, is believed to be the northern extension of the Clarendon-Linden fault system (McFall, 1993). Farther north, the projection of the Clarendon-Linden fault system and Salmon River faults coincide with the Robertson Lake mylonite zone in the Canadian Shield (McFall, 1993; Easton and others, 1995).

The Clarendon-Linden fault system probably represents brittle reactivation of a major Grenville structure (Hutchinson and others, 1979; Seeber and Armbruster, 1995). The detailed map pattern of the Clarendon-Linden fault system shows that the north-south-trending faults comprise numerous short segments connected by oblique, northwest-trending transfer faults,

similar to that in rift settings (Jacobi and Fountain, 1996). This geometry suggests that the underlying east-dipping Precambrian thrusts could have been acting as detachment, possibly during Iapetan rifting (Jacobi and Fountain, 1996). Culotta and others (1990) have correlated the Clarendon-Linden fault system with a continental scale magnetic lineament, the Amish anomaly, which extends from New York to Alabama (Culotta and others, 1990). This feature is interpreted as a major Grenville terrane boundary, separating the Elzevir and Frontenac tectonic blocks (Culotta and others, 1990). Wheeler (1995) has placed the northwestern boundary of the Iapetan rifted margin along the Clarendon-Linden fault system, which implies it is a potentially seismogenic structure.

Seismogenic Potential. The following table summarizes the assessment of the seismogenic potential of the Clarendon-Linden fault system. The individual assessments are described below.

Criterion (W)	Probability (P)
Association with $m_b \geq 5$ seismicity	0.1
Association with $2 \leq m_b < 5$ seismicity	0.25
Evidence for reactivation	1.0
Slip favorable w/ present stress regime	0.1
Seismogenic crustal extent	0.6
Probability of being seismogenic	0.25

The Clarendon-Linden fault system is associated with a well-defined cluster of seismicity, which is referred to as the Attica seismic zone (Seeber and Armbruster, 1995). This zone contains the 1929 m_b 5.2 Attica earthquake, numerous natural small-magnitude events, and several induced events associated with salt brine recovery at Dale, New York (Fletcher and Sykes, 1977). Seeber and Armbruster (1995) suggest it also is possible that the 1929 event may have been artificially triggered, as the brine fields in Dale were already active in 1929. Seismicity appears to be spatially associated, and possibly aligned with the gravity gradient that marks the approximate location of the Clarendon-Linden fault system. However, while

seismicity does appear to be spatially associated with a segment of the Clarendon-Linden fault system, it does not form an alignment along a significant portion of the structure. Given the observation that the 1929 Attica earthquake occurred within 50 km of the Clarendon-Linden fault system and is the only known $m_b \geq 5$ event in the area, a low probability (0.1) is assigned to the Clarendon-Linden fault system being associated with $m_b \geq 5$ seismicity.

Based on the cluster of small-magnitude seismicity that is spatially associated and partially aligned with a short segment of the Clarendon-Linden fault system, but does not form a continuous alignment along the length of the feature, a relatively low probability (0.25) is assigned to the Clarendon-Linden fault system being associated with small-magnitude seismicity.

Deep seismic reflection profiles and geophysical data suggest that the Clarendon-Linden fault system is associated with basement structures that extend to seismogenic depths (Forsyth and others, 1994b; Culotta and others, 1990). Although these basement structures cannot be directly linked to individual Clarendon-Linden faults, the available evidence suggests that a structural linkage between the features is likely. We therefore assign a moderate probability (0.6) to the Clarendon-Linden fault system extending to seismogenic depths.

There is no clear evidence of brittle slip on the Clarendon-Linden fault system during the present tectonic stress regime. Jacobi and Fountain (1993) reported that gas seeps along the southern part of Clarendon-Linden fault system were initiated during, or shortly after the 1988 Saguenay (m_{bLg} 6.5) earthquake. The source of the gas seeps was inferred to be a Devonian shale, at a depth of about 300 m (Jacobi and Fountain, 1993). Jacobi and Fountain (1993) suggested that fractures within the fault zone were opened as a local response to the Saguenay event. While the opening of fractures and the emission of deep gases from the Clarendon-Linden fault system suggests recent activity, there is no evidence to suggest that this activity was accompanied by brittle tectonic displacement at depth. There is no evidence of recent tectonic deformation at the ground surface. Based on the foregoing information, a very low probability (0.1) is assigned to the Clarendon-Linden fault system having evidence of brittle slip in the present day regional tectonic stress regime.

Well data indicate that the Clarendon-Linden fault system was reactivated twice during the Paleozoic, once in extension and once in compression (Van Tyne, 1975). From early to middle Ordovician time the Clarendon-Linden fault system was periodically active as normal fault, while its post-Devonian sense of displacement has been reverse (Van Tyne, 1975; Jacobi and Fountain, 1993). Jacobi and Fountain (1996) also have presented seismic reflection data that indicate multiple periods of Paleozoic reactivation of the Clarendon-Linden fault system. Based on this information, a very high probability (1.0) is assigned to the Clarendon-Linden fault system having evidence of multiple episodes of reactivation.

Based on our assessment, we arrive at the probability of 0.25 that the Clarendon-Linden fault system is seismogenic.

Additional Information. Additional information exists regarding the seismogenic potential of the Clarendon-Linden fault zone. Herrmann (1978) reported that focal mechanism studies of two shallow (2-3 km) earthquakes in 1966 and 1967 in the Attica area suggested seismicity occurs along approximately north-northeast-trending Clarendon-Linden faults. The 1966 event (m_b=4.6) yielded solutions with one nodal plane striking NNE and dipping steeply to the east, and a second striking WNW and dipping steeply to the south. If the NNE solution is accepted, then the fault motion was primarily right-lateral strike-slip with a reverse component. The 1967 event (m_b=4.4) yielded solutions with one nodal plane striking NNE and dipping steeply to the east, and a second WNW striking plane dipping moderately to the south. Again, accepting the NNE solution indicates right-lateral and reverse slip. Herrmann (1978) selected the NNE nodal plane as the most likely fault plane solution, in light of the well log data (Van Tyne, 1975) indicating NNE-trending Clarendon-Linden faults. He noted, however, that the WNW trend could not be totally discounted.

The focal mechanisms derived from these two events yield solutions that are consistent with the geometry (both strike and dip) of the Clarendon-Linden faults as determined from both geologic and seismic data. We judge that it is twice as likely that a consistent focal mechanism will be observed for a given seismogenic feature compared to a given nonseismogenic feature (relative likelihood 2:1).

Considering this additional information and using Equation (3-2), the probability that the Clarendon-Linden fault system is seismogenic is updated from 0.25 to 0.40.

The Georgian Bay Linear Zone

The Georgian Bay linear zone was identified by Wallach and Mohajer (1990) on the basis of parallel, discontinuous linear aeromagnetic and gravity anomalies. The feature comprises an approximate 50 km-wide zone extending from the northeast corner of Georgian Bay to western New York State, coincident with the straight east coast of Georgian Bay (Wallach and Mohajer, 1990) (Figure 2-12). Wallach (1990) interpreted the differently expressed lineaments of the Georgian Bay linear zone as a major Precambrian structural zone. A physiographic lineament along the trend of the Georgian Bay linear zone was interpreted by Sanford and others (1985) as a fault block boundary, tectonically active at least during the early Paleozoic. The Laurentian channel, a pre-glacial outlet of the upper Great Lakes to the St. Lawrence River (Spencer, 1907; Eyles and others, 1993), also lies along the trend of the Georgian Bay linear zone.

Based on interpretations of aeromagnetic and gravity data in southern Ontario, Roest (1995) reported the existence of the Georgian Bay linear zone to be very questionable. He stated that "the Georgian Bay linear zone to the south of Georgian Bay proper is not present in horizontal gravity gradient maps, and is questionable in aeromagnetic data, due to flight line orientation and survey edge artifacts." Roest (1995) also noted that the relatively low gravity gradient associated with the Georgian Bay linear zone is not indicative of a fault. For comparison, the Clarendon-Linden fault zone is associated with relatively large gravity gradients of >1.5 mGal/km (Roest, 1995).

Wallach and Mohajer (1990) reported that a zone of earthquakes trending north-northwest extends from western New York State to near the northeast corner of Georgian Bay and outlines a linear belt which lies within, and parallels the elements of the Georgian Bay linear zone. Among these earthquakes are the 1857 Lockport event (~M 5.0) (since relocated southwest of Buffalo and downgraded to M~4) and the 1929 Attica earthquake (m_b 5.2), the Attica earthquake occurring near the intersection of the Georgian Bay linear zone and the

Clarendon-Linden fault system (Wallach, 1995a). Two mesoscopic brittle faults observed in lower Paleozoic rocks in the vicinity of the Georgian Bay linear zone are thought to be associated with this feature (Wallach and Mohajer, 1990; Wallach, 1995a). These faults are oriented approximately parallel to the zone, with one exhibiting 5 cm of normal displacement and dipping 75° northeast, and the other dipping 75° southwest with a minimum of 5.5 m of normal displacement (Wallach and Mohajer, 1990). Wallach (1995a) also has suggested that the northwest-striking normal faults identified along the Rouge River (e.g., Mohajer and others, 1992; Hibbert and others, 1994) may be associated with this feature. Wallach (1995a) reported that the presence of brittle structures and the apparent zone of earthquakes along the trend of the Georgian Bay linear zone implies that the zone may be seismically active.

Seismogenic Potential. The Georgian Bay linear zone is potentially associated with the Rouge River fault zone (Wallach, 1995a), as discussed in Section 4, and this assessment has an effect on the probability of the feature being seismogenic. Following the logic tree shown in Figure 4-2, two assessments are made for the Georgian Bay linear zone; one considers an association with the Rouge River faults, the other does not. We consider an association of these features based on the observations that they have similar trends, and the Rouge River faults lie within the Georgian Bay linear zone. The following tables summarize the assessment of the seismogenic potential of the Georgian Bay linear zone. The individual assessments are discussed below. Unless otherwise stated, the assessment is considered for both cases, i.e., an association with the Rouge River faults, and no association with the Rouge River faults.

Assumed association with the Rouge River faults:

CRITERION (W)	PROBABILITY (P)
Association with $m_b \geq 5$ seismicity	0.1
Association with $2 \leq m_b < 5$ seismicity	0.1
Evidence for reactivation	0.2
Slip favorable w/ present stress regime	0.2
Seismogenic crustal extent	0.1
Probability of being seismogenic	0.13

No association with the Rouge River faults:

CRITERION (W)	PROBABILITY (P)
Association with $m_b \geq 5$ seismicity	0.1
Association with $2 \leq m_b < 5$ seismicity	0.1
Evidence for reactivation	0.1
Slip favorable w/ present stress regime	0.1
Seismogenic crustal extent	0.1
Probability of being seismogenic	0.12

The 1929 m_b 5.2 Attica, New York earthquake is located at the southeast end of the Georgian Bay linear zone, and most interpretations associate this event with the Clarendon-Linden fault system (Herrmann, 1978; Seeber and Armbruster, 1993; Tuttle and others, 1996). However, given the error (<50 km) associated with the epicentral location of the Attica event, it could be associated with the Georgian Bay linear zone. There is also uncertainty in the epicentral location and magnitude of the 1857 Lockport event, however, the most recent interpretation revises its magnitude downward and places it southwest of Buffalo; its previous location placed it within the Georgian Bay linear zone. Given the possibility of a spatial association with the 1929 Attica event, we assign a very low probability (0.1) to the Georgian Bay linear

zone being associated with $m_b \geq 5$ seismicity. In light of the recent re-interpretation of the 1857 Lockport event, we do not consider it to be spatially associated with this zone.

Wallach and Mohajer (1990) reported that at least eight small-magnitude earthquakes, both historical and instrumentally recorded events, define a line which extends along the Georgian Bay linear zone from northwestern New York State to the northeastern corner of Georgian Bay. Several small-magnitude events are located within the mapped trace of the Georgian Bay linear zone in the Lake Ontario vicinity, and numerous others occur adjacent and proximal to it in the northeast. The northwest-trending gravity gradient south of Lake Ontario near the Clarendon-Linden fault system appears to be spatially associated with some seismicity. However, the small-magnitude seismicity that is spatially associated with the Georgian Bay linear zone occurs as small, isolated clusters in parts of the zone, does not define a continuous alignment along it, and consists of relatively few events such that the number of events within the zone is not greatly different from that occurring outside of it. Given the rather weak spatial association of some events with this feature and the absence of a linear alignment along it, a very low probability (0.1) is assigned to the Georgian Bay linear zone being associated with small-magnitude seismicity.

There are no data to directly evaluate the downdip extent of the Georgian Bay linear zone. Roest (1995) reported that the very existence of this feature in horizontal gravity gradient and aeromagnetic maps is questionable, and that the associated gravity gradient is generally low and not indicative of a fault. Furthermore, unlike the Niagara Pickering linear zone and the Clarendon-Linden fault system, Rankin and others (1993) do not define a major tectonic boundary that is spatially coincident with this feature. The relatively weak, discontinuous geophysical expression of the Georgian Bay linear zone suggests it is not a prominent crustal structure. The northwest-trending transfer zones among the north-south-striking Clarendon-Linden faults could be considered analogous to hypothesized features comprising the Georgian Bay linear zone. However, these northwest-trending structures occur in the Paleozoic cover, and do not unequivocally extend to seismogenic depths. Given the weak geophysical expression of this feature relative to the other geophysical lineaments in the

region, and its discontinuous nature, we assign a low probability (0.1) to the Georgian Bay linear zone extending to seismogenic depths.

Wallach (1995a) reported that a northwest-trending linear feature, several hundred meters long was recognized in the glaciolacustrine sediments of Georgian Bay (Blasco, personal communication cited in Wallach, 1995a). This feature occurs parallel to the Georgian Bay linear zone, and was interpreted as resulting from degassing or dewatering (Blasco, personal communication cited in Wallach, 1995a), which suggests the presence of an open fracture or active fault (Thomas, personal communication cited in Wallach, 1995a). While the orientation of this feature is considered favorable for reactivation in the current tectonic stress regime, there is no evidence that the feature is accompanied by brittle tectonic displacement. Given the uncertainty in the origin of this feature, and the ambiguous association of this feature with the Georgian Bay linear zone, we assign a very low probability (0.1) to the Georgian Bay linear zone having evidence for brittle slip consistent with the contemporary tectonic stress regime. If an association between the Georgian Bay linear zone and the Rouge River faults is assumed, we assign a low probability of (0.2) to this criterion. This assessment considers the evidence discussed above, and reflects the observations that the Rouge River faults are Quaternary structures favorably oriented with respect to the regional tectonic stress field, but their normal sense of displacement is not consistent with northeast-southwest oriented compression.

Wallach and Mohajer (1990) reported evidence of brittle deformation in both the Precambrian and Paleozoic rocks within the Georgian Bay linear zone. Two small-displacement northwest-striking normal faults, whose ages are unknown, occur in lower Paleozoic rocks along the trend of the Georgian Bay linear zone and are postulated to be associated with it (Wallach and Mohajer, 1990). There is no evidence indicating a structural association of these local, outcrop-scale faults with the postulated basement structures underlying the Georgian Bay linear zone, other than the similarity in their trends. Small displacement faults occur in the lower Paleozoic rocks throughout southern Ontario, and these structures have not been studied systematically to determine if they are anomalous in the vicinity of the Georgian Bay linear zone.

Well log analysis by Pearce (1991) in upper New York State suggested to him that a northwest-trending feature controlled deposition in the Silurian salt section, implying it was active during the Silurian. The northwest-trending fault zones associated with the Clarendon-Linden fault system were active in the Ordovician, Devonian and post-Devonian (Jacobi and Fountain, 1996), and it is possible that these faults could be considered analogous to hypothesized structures in the Georgian Bay linear zone. Eyles and others (1993) suggested that the position of the Laurentian channel, a northwest-trending pre-glacial outlet between Georgian Bay and Lake Ontario, is controlled by a basement structure, i.e., the Georgian Bay linear zone. The foregoing information is considered as possible evidence of reactivation of hypothesized structures comprising the Georgian Bay linear zone. However, given the geophysical evidence suggesting that the Georgian Bay linear zone is not a Grenville crustal structure (e.g., Roest, 1995), we assign a very low probability (0.1) to the Georgian Bay linear zone having multiple episodes of reactivation. If an association between the Georgian Bay linear zone and the Rouge River faults is assumed, we assign a low probability of (0.2) to this criterion. This assessment reflects the evidence discussed above, along with the observations that: 1) these Quaternary faults have a similar geometry to the two mesoscopic faults identified by Wallach and Mohajer (1990), 2) the involvement of bedrock faulting along the Rouge River is equivocal, and 3) normal faulting is not consistent with the present compressional tectonic stress regime.

Considering the association of the Rouge River faults with the Georgian Bay linear zone, we arrive at the probability of 0.13 that the Georgian Bay linear zone is seismogenic. When the Rouge River faults are assumed not to be associated with the Georgian Bay linear zone, we arrive at a probability of 0.12 that the Georgian Bay linear zone is seismogenic. The combined probability that the Rouge River faults are associated with the Georgian Bay linear zone, and the Georgian Bay linear zone is seismogenic is (0.15 x 0.13 = 0.02). Likewise, the combined probability that the Rouge River faults are *not* associated with the Georgian Bay linear zone, and the Georgian Bay linear zone is seismogenic is (0.85 x 0.12 = 0.102). (*See* Section 4 for a discussion). As discussed above, the observation of Quaternary faulting at Rouge River is not considered to provide additional information to update the probability of the feature being seismogenic.

The Hamilton-Presqu'ile Fault

The Hamilton-Presqu'ile fault strikes east-northeast and extends for approximately 200 km. from near the western shore of Lake Ontario to Prince Edward County in the northeast (Ontario Geological Survey, 1991) (Figure 3-1). This feature was recognized by McFall and Allam (1991) as a discontinuous aeromagnetic lineament in the Precambrian basement beneath Lake Ontario, and was later classified as a subsurface fault by the Ontario Geological Survey (1991). Early Paleozoic rocks exposed in Prince Edward county are displaced 40 m down to the north across the eastern end of the Hamilton-Presqu'ile fault (LeBaron and Williams, 1990). McQuest Marine (1995) reported that four faults, displacing lake bottom sediments estimated to be about 11 ka and bedrock, can be seen in airgun seismic reflection profiles collected across this feature in the vicinity of the intersection of the Hamilton-Presqu'ile fault with the Niagara-Pickering linear zone. The interpreted faults are northeast-striking and predominantly north-dipping, and have apparent throws of up to 7 m (Figure 3-11). Sanford (1995) also interpreted numerous small-displacement northeast-trending fractures affecting the Paleozoic bedrock and post-Paleozoic sediments in this area from sleeve airgun profiles. Lewis and others (1995) interpreted several small displacement (0.5 m) normal faults in the same general area of western Lake Ontario. The interpreted faults occur in acoustically stratified sediment also estimated to be about 11 ka, and Lewis and others (1995) suggested that differential settling of the underlying sediment could be responsible for the features. Lewis and others (1995) also documented numerous sediment depressions in this area, and the margins of some of these depressions resemble monoclines. These features were attributed to sediment drape into a depression, rather than displacement of sediment after deposition, based on continuous and undisturbed reflection data (Lewis and others, 1995). The depressions could not be correlated to deeper sedimentological or bedrock features, which strongly suggests a nontectonic origin (Lewis and others, 1995).

Thomas and others (1993) noted that the Hamilton-Presqu'ile fault lies parallel to, and just north of, the northern margin of the postulated southwest extension of the St. Lawrence rift system as illustrated by Adams and Basham (1989). Thomas and others (1993) thus interpreted this structure to represent either the northern boundary of the St. Lawrence rift extension, or one of a series of faults within a rift zone wider than previously envisaged. It

should be noted that, at the intersection of the Hamilton-Presqu'ile fault with the Niagara-Pickering linear zone, the east-west-trending magnetic lineament associated with the Hamilton-Presqu'ile fault appears to be truncated by the north-northeast-trending Niagara-Pickering magnetic lineament.

Seismogenic Potential. The following table summarizes the assessment of the seismogenic potential of the Hamilton-Presqu'ile fault. The individual assessments are characterized below.

CRITERION (W)	PROBABILITY (P)
Association with $m_b \geq 5$ seismicity	0
Association with $2 \leq m_b < 5$ seismicity	0.15
Evidence for reactivation	0.2
Slip favorable w/ present stress regime	0.1
Seismogenic crustal extent	0.5
Probability of being seismogenic	0.08

Both of the known $m_b \geq 5$ earthquakes in the region occur well over 50 km away from the Hamilton-Presqu'ile fault. Given that the error in the epicentral locations of these events is less than 50 km, we assign a probability of zero to the Hamilton-Presqu'ile fault being associated with $m_b \geq 5$ seismicity.

Mohajer (1993) reported that at least ten small-magnitude earthquakes (historical and instrumentally recorded events) are possibly associated with the Hamilton-Presqu'ile fault. Seven of these earthquakes occur at, or near, the intersection of the Toronto-Hamilton seismic zone and the Hamilton-Presqu'ile fault, and three of the events can be associated with either feature (Mohajer, 1993). Seismicity that could be spatially associated with the Hamilton-Presqu'ile fault is relatively scattered, and does not form an alignment along the feature. A weak cluster occurs near the eastern end of the Hamilton-Presqu'ile fault, which is proximal to the Wilson-Port Hope lineament and could be associated with either structure. Given the scarcity of events proximal to this feature and the absence of an alignment of seismicity along

it, we assign a very low probability (0.15) to the Hamilton-Presqu'ile fault being associated with small-magnitude seismicity.

There are no data to directly evaluate the downdip geometry and extent of the Hamilton-Presqu'ile fault. However, based on its mapped length (~200 km) and the 40 m of displaced Paleozoic rocks across this feature, a moderate probability (0.5) is assigned to the Hamilton-Presqu'ile fault extending to seismogenic depths.

Evidence for brittle slip in the present stress regime consists of the displaced Paleozoic rocks across the eastern end of the fault and the interpreted faults on the lake bottom (e.g., McQuest Marine, 1995; Sanford, 1995). The interpreted seismic reflection data of McQuest Marine (1995) is shown in Figure 3-11. From the figure it appears that the interpreted faults exhibit a downward decrease in displacement, which suggests they are not tectonic structures. Further, the vertical exaggeration of the cross-section is over 60 times, which grossly distorts the actual relief of the reflectors. Sanford (1995) interpreted numerous apparent block faults affecting bedrock and overlying post-Paleozoic sediments along the trend of the Hamilton-Presqu'ile fault in western Lake Ontario. However, the timing and sense of movement on these structures could not be precisely established (Sanford, 1995). Although the northeast orientation of the faults interpreted by McQuest Marine (1995) is considered favorable for reactivation in the present tectonic stress field, their normal sense of displacement is not consistent with northeast-southwest oriented compression. Further, their apparent downdip decrease in offset strongly suggests a nontectonic origin. The normal sense of displacement of the Paleozoic rocks across the east end of this fault also is considered inconsistent with regional northeast-oriented compression. The origin, age, and sense of displacement of the features interpreted by Sanford (1995) is questionable, however, their orientations suggest a possible association with the Hamilton-Presqu'ile fault. Based on the available evidence, we assign a very low probability (0.1) to the Hamilton-Presqu'ile fault having evidence of brittle faulting compatible with the present day regional tectonic stress regime.

McFall and Allam (1990) originally defined this structure on the basis of aeromagnetic data, which implies that it lies within the Grenville basement since the overlying Paleozoic cover

has relatively no magnetic signal (McFall, 1993). Ordovician rocks displaced 40 m down to the north across the eastern end of the Hamilton-Presqu'ile fault are considered as potential evidence for an episode of reactivation of this structure. Additional evidence of brittle reactivation of the Hamilton-Presqu'ile fault consists of the interpreted lake bottom faults along the trend of the Hamilton-Presqu'ile fault in western Lake Ontario. Sanford (1995) interpreted block faulting of the Ordovician rocks in this area, and he suggested that some of these faults may have moved in post-Paleozoic time. However, it is unknown if these structures are associated with the Hamilton-Presqu'ile fault. None of the faults interpreted by Sanford (1995) can unequivocally be shown to displace the recent lake bottom sediments. Based on this information, a low probability (0.2) is assigned to the Hamilton-Presqu'ile fault having evidence of multiple episodes of reactivation.

Based on the above assessment, we arrive at a probability of 0.08 that the Hamilton-Presqu'ile fault is seismogenic.

Rouge River Faults

Normal faults that offset Quaternary deposits in the Rouge River Valley east of metropolitan Toronto have been interpreted by Mohajer and others (1992, 1995) as neotectonic features that offset the underlying bedrock. Adams and others (1993), however, interpreted the structures to be the result of glacier overriding, or glacial ice shove (glaciotectonism). The faults strike west-northwest, and the maximum vertical displacement observed in outcrop is 1.25 m (Mohajer and others, 1992) (Figure 3-12). Mohajer (1995a) reported that the age of the faulting is poorly constrained, but is believed to be no older than 40 ka, and may postdate 13 ka. Adams and others (1993) interpreted the structures as predating 13 ka, and representing one episode of displacement in about 125 ka. Wallach (cited in Mohajer, 1995a) suggested that at least two episodes of normal faulting have occurred within the past 40,000 years.

Mohajer and others (1992) initially interpreted the structures to be the result of glacial rebound aided by high horizontal compressive stress. Mohajer (1995a) later suggested that the faults represent reactivation of Grenville basement structures. Adams and others (1993), however, noted that the effects of glaciotectonism are visible nearby (within 50 m) at the same

stratigraphic level, and that normal faulting is not uncommon in glaciotectonized sedimentary sequences (e.g. van der Wateren, 1987; Broster, 1991). They also noted that some of the faults have downward-decreasing offset, which is inconsistent with a tectonic origin, and others have a constant throw. Adams and others (1993) further suggested that the extreme Pleistocene glaciation in the Toronto area supports a glacial mechanism for the formation of the faults.

Hibbert and others (1994) identified additional Quaternary faulting in the Rouge River Valley upstream from the area investigated by Mohajer and others (1992). Multiple normal faults were discovered, having an apparent throws of up to 4 m (Hibbert and others, 1994). These faults also have west-northwest strikes, and displace Quaternary deposits estimated to be younger than the faulted Quaternary deposits investigated by Mohajer and others (1992). Hibbert and others (1994) interpreted repeated episodes of deformation, based on the number faults present, and additional evidence for slumping and re-sedimentation across the faults.

Mohajer and others (1995) collected more than 1.5 km of high-resolution seismic reflection data across the fault zones mapped in the Rouge River. They interpreted a profile collected across the faults investigated by Hibbert and others (1994) as revealing a maximum offset of 15 m of the bedrock surface and an offset of about 4 m of the overlying Quaternary strata (Figure 3-13). Mohajer and others (1995) suggested that this indicates an upwards decrease in displacement along the fault, which supports a tectonic origin for the Rouge River faults, as opposed to the glacial mechanism proposed by Adams and others (1993).

Seismogenic Potential. The assessment of the seismogenic potential of the Rouge River fault is a function of whether the fault is considered to be a separate seismic source or part of other potential sources such as the Niagara-Pickering linear zone or the Georgian Bay linear zone. The logic structure for considering the fault is discussed in detail in Section 4. Here we consider the fault as a separate potential seismic source.

The following table summarizes our assessment of the seismogenic potential of the Rouge River faults. The individual assessments are characterized below.

CRITERION (W)	PROBABILITY (P)
Association with $m_b \geq 5$ seismicity	0
Association with $2 \leq m_b < 5$ seismicity	0.1
Evidence for reactivation	0.1
Slip favorable w/ present stress regime	0.1
Seismogenic crustal extent	0.1
Probability of being seismogenic	0.04

Both of the known $m_b \geq 5$ earthquakes in the region occur over 100 km from the Rouge River faults. The errors in the epicentral locations of these events are less then 50 km, and we therefore assign a probability of zero to the Rouge River faults being associated with $m_b \geq 5$ seismicity.

Although there is a concentration of small-magnitude seismicity in the western Lake Ontario region, few events are located within a 50 km radius of the Rouge River faults. There is no obvious spatial association of small-magnitude earthquakes with these structures. However, we assign a very low probability (0.1) to this criteria to allow for the possibility that an event within 50 km of the features could be associated.

There are no data to directly evaluate the downdip extent of the Rouge River faults. Shallow seismic reflection data collected across the faults by Mohajer and others (1995) do not, because of their very shallow nature, provide any evidence to suggest the faults extend to significant depths. Mohajer and others (1995) interpreted bedrock faulting from the shallow seismic data based on an apparent step at the bedrock surface (Figure 3-13). However, the seismic reflection data do not show unequivocally that faults extend through the bedrock. Furthermore, Broster (1991) has documented normal faults in bedrock that formed due to glacial overriding. In view of the observations supporting a nontectonic origin of these

features (e.g., Adams and others, 1993), and their relatively small dimensions, a very low probability (0.1) is assigned to the Rouge River faults extending to seismogenic depths.

The west-northwest-striking Rouge River faults are considered to be favorably oriented for tectonic movement with respect to the present day stress regime. Their normal sense of displacement, however, is not consistent with the contemporary northeast-southwest oriented compressional stress field. Furthermore, focal mechanisms from earthquakes in the region indicate the occurrence of mainly thrust faulting events (Adams and others, 1991). However, allowing for the possibility that these extensional features are associated with hanging-wall deformation in an (as yet undetected) underlying blind reverse fault, a very low probability (0.1) is assigned to the Rouge River faults having brittle faulting consistent with the present day regional stress regime.

There is no direct evidence for multiple tectonic episodes of reactivation on the Rouge River faults. The 15 m of interpreted bedrock-offset suggests multiple faulting events, however, there is no evidence to suggest that faulting occurred during distinctly different tectonic phases. Mohajer (1995) and Wallach (1995) have suggested that the Rouge River faults result from reactivated basement structures. Based on the available data, we assign a very low probability (0.1) to the Rough River faults having evidence for multiple episodes of tectonic reactivation.

Based on our assessment, we arrive at a probability of 0.05 that the Rouge River faults are seismogenic. As discussed above, the observation of Quaternary faulting at Rouge River is not considered to provide additional information to update the probability of the feature being seismogenic.

Darlington Fracture Zone

Ontario Hydro (1977) identified a set of approximately east-west-trending, steeply south-dipping joints exposed in the St. Mary's limestone quarry, which is located less than 5 km to the northeast of the Darlington site, during a geologic hazards investigation conducted for the Darlington Generating Station. Subsequent mapping of bedrock excavations at the Darlington

site (Ontario Hydro, 1981) confirmed the presence of the approximately east-west trending joints at the site, and revealed subhorizontal joints parallel to bedding, and less prominent joints trending roughly north-south. The joints were described as being generally straight, having smooth to slightly rough surfaces with no slickenslides, and the observed joint spacing ranged from less than 0.3 m up to 2.5 m (Ontario Hydro, 1981). No evidence of faulting was observed in the bedrock exposures (Ontario Hydro, 1977, 1981).

Side-scan sonar surveys of the region offshore of the Darlington site were conducted by Ontario Hydro in 1987 and 1988. These surveys revealed the presence of a zone of anomalous north-northeast-trending linear features in bedrock, which were interpreted by a consultant as a major fault zone (Ontario Hydro, 1989). Subsequent seismic reflection profiling across these features produced a series of unbroken, nearly horizontal reflections that could be followed to a bedrock outcrop on the lakebed. It was concluded by Earth & Ocean Research, Ltd. (Ontario Hydro, 1989) that the anomalous linear features were related to differential, erosional terracing of fine scale bedding within the outcropping bedrock, and not faulting. No evidence of faulting was observed during the study (Ontario Hydro, 1989).

Wallach (1995c) suggested that a northeast-trending linear break in slope evident on bathymetric and bedrock surface maps produced by Ontario Hydro (1989) might indicate the presence of a fault. This bathymetric feature trends subparallel to the anomalous features identified by Earth & Ocean Research, Ltd. (Ontario Hydro, 1989), and a linear positive magnetic anomaly interpreted as a fault by a consultant (Patterson, Grant and Watson, cited in Wallach, 1995c). Submersible diving in 1994 confirmed the presence of the bathymetric lineament, and although murky waters prevented drawing any conclusions regarding the feature, a normal stratigraphic sequence (Lindsay Formation overlain by Whitby Formation) was observed (Wallach, 1995c). Wallach (1995c) reported that the elevation of the observed geologic contact in the vicinity of the break in slope appears to be anomalous, and may or may not be due to faulting. He further stated that no firm conclusions could be drawn from this observation, and it remains unclear whether or not the break in slope represents a fault, or an artifact of something else. In view of the lack of evidence suggesting that this bathymetric feature is a fault, we do not make an assessment for its seismogenic potential.

TABLE 3-1

CONDITIONAL PROBABILITIES OF BEING SEISMOGENIC AS A FUNCTION OF OBSERVED CRITERIA

				Multiple Reactivations							
				Yes				No			
				Brittle Slip				Brittle Slip			
				Yes		No		Yes		No	
				Deep Crustal Extent		Deep Crustal Extent		Deep Crustal Extent		Deep Crustal Extent	
				Yes	No	Yes	No	Yes	No	Yes	No
Spatial Association $m_b \geq 5$	Yes	Spatial Association $2 \leq m_b < 5$	Yes	0.95	0.94	0.93	0.92	0.91	0.90	0.87	0.85
			No	0.90	0.89	0.88	0.88	0.86	0.85	0.82	0.80
	No	Spatial Association $2 \leq m_b < 5$	Yes	0.70	0.68	0.57	0.55	0.52	0.50	0.42	0.40
			No	0.10	0.08	0.05	0.02	0.01	0.005	0.002	<0.001

Table 3-2
Example Calculation of the Probability a Feature is Seismogenic

Assoc $m_b \geq 5$		Assoc $m_b < 5$		Reactivation		Brittle Slip		Deep Crustal Extent		Comb. Prob of State*	Cond. Prob. of Seis.–	Prob. of Seis.=–
State	Prob.	State	Prob.	State	Prob.	State	Prob.	State	Prob.			
Yes	0.1	Yes	0.4	Yes	0.3	Yes	0.6	Yes	0.7	0.0050	0.95	0.0048
Yes	0.1	Yes	0.4	Yes	0.3	Yes	0.6	No	0.3	0.0022	0.94	0.0021
Yes	0.1	Yes	0.4	Yes	0.3	No	0.4	Yes	0.7	0.0034	0.93	0.0032
Yes	0.1	Yes	0.4	Yes	0.3	No	0.4	No	0.3	0.0014	0.92	0.0013
Yes	0.1	Yes	0.4	No	0.7	Yes	0.6	Yes	0.7	0.0118	0.91	0.0107
Yes	0.1	Yes	0.4	No	0.7	Yes	0.6	No	0.3	0.0050	0.90	0.0045
Yes	0.1	Yes	0.4	No	0.7	No	0.4	Yes	0.7	0.0078	0.87	0.0068
Yes	0.1	Yes	0.4	No	0.7	No	0.4	No	0.3	0.0034	0.85	0.0029
Yes	0.1	No	0.6	Yes	0.3	Yes	0.6	Yes	0.7	0.0076	0.90	0.0068
Yes	0.1	No	0.6	Yes	0.3	Yes	0.6	No	0.3	0.0032	0.89	0.0028
Yes	0.1	No	0.6	Yes	0.3	No	0.4	Yes	0.7	0.0050	0.88	0.0044
Yes	0.1	No	0.6	Yes	0.3	No	0.4	No	0.3	0.0022	0.87	0.0019
Yes	0.1	No	0.6	No	0.7	Yes	0.6	Yes	0.7	0.0176	0.86	0.0151
Yes	0.1	No	0.6	No	0.7	Yes	0.6	No	0.3	0.0076	0.85	0.0065
Yes	0.1	No	0.6	No	0.7	No	0.4	Yes	0.7	0.0118	0.82	0.0097
Yes	0.1	No	0.6	No	0.7	No	0.4	No	0.3	0.0050	0.80	0.0040
No	0.9	Yes	0.4	Yes	0.3	Yes	0.6	Yes	0.7	0.0454	0.70	0.0318
No	0.9	Yes	0.4	Yes	0.3	Yes	0.6	No	0.3	0.0194	0.68	0.0132
No	0.9	Yes	0.4	Yes	0.3	No	0.4	Yes	0.7	0.0302	0.57	0.0172
No	0.9	Yes	0.4	Yes	0.3	No	0.4	No	0.3	0.0130	0.55	0.0072
No	0.9	Yes	0.4	No	0.7	Yes	0.6	Yes	0.7	0.1058	0.52	0.0550
No	0.9	Yes	0.4	No	0.7	Yes	0.6	No	0.3	0.0454	0.50	0.0227
No	0.9	Yes	0.4	No	0.7	No	0.4	Yes	0.7	0.0706	0.42	0.0297
No	0.9	Yes	0.4	No	0.7	No	0.4	No	0.3	0.0302	0.40	0.0121
No	0.9	No	0.6	Yes	0.3	Yes	0.6	Yes	0.7	0.0680	0.10	0.0068
No	0.9	No	0.6	Yes	0.3	Yes	0.6	No	0.3	0.0292	0.08	0.0023
No	0.9	No	0.6	Yes	0.3	No	0.4	Yes	0.7	0.0454	0.05	0.0023
No	0.9	No	0.6	Yes	0.3	No	0.4	No	0.3	0.0194	0.02	0.0004
No	0.9	No	0.6	No	0.7	Yes	0.6	Yes	0.7	0.1588	0.01	0.0016
No	0.9	No	0.6	No	0.7	Yes	0.6	No	0.3	0.0680	0.005	0.0003
No	0.9	No	0.6	No	0.7	No	0.4	Yes	0.7	0.1058	0.002	0.0002
No	0.9	No	0.6	No	0.7	No	0.4	No	0.3	0.0454	0.001	0.0000
									Sum	1.0000	Sum	0.2903

* Probability of combined state of criteria equal to product of probabilities for each criteria.
– Conditional probability of being seismogenic from Table 3-1.
= Product of probability of combined state and conditional probability of being seismogenic. Sum is the unconditional probability of being seismogenic.

TABLE 3-3
SUMMARY OF ASSESSMENTS AND THE SEISMOGENIC POTENTIAL OF LOCAL SOURCES

	Spatial Association with $m_b \geq 5$ Seismicity	Spatial Association with m_b 2-5 Seismicity	Geologic Evidence for Multiple Episodes of Reactivation	Geologic Evidence for Brittle Slip in Present Stress/Tectonic Regime	Extends to Seismogenic Depths (10-20 km)	Probability of Being Seismogenic P(S)	Updated P(S) (Considering Additional Information)
Toronto-Hamilton Seismic Zone	0	0.3	0.4	0.1	0.9	0.16	
Niagara-Pickering Linear Zone							
w/ RRF* and w/o AMB**	0	0.1	0.6	0.3	1	0.09	
w/o RRF and w/o AMB	0	0.1	0.5	0.2	1	0.08	
w/ AMB and RRF	0.2	0.25	0.6	0.3	1	0.31	0.36
w/ AMB and w/o RRF	0.2	0.25	0.5	0.2	1	0.30	0.35
Wilson-Port Hope Lineament	0	0.1	0.4	0.1	0.2	0.06	
Clarendon-Linden Fault System	0.1	0.25	1	0.1	0.6	0.25	0.4
Georgian Bay Linear Zone							
w/ RRF	0.1	0.1	0.2	0.2	0.1	0.13	
w/o RRF	0.1	0.1	0.1	0.1	0.1	0.12	
Hamilton-Presqu'ile Fault	0	0.15	0.2	0.1	0.5	0.08	
Rouge River Faults	0	0.1	0.1	0.1	0.1	0.04	
* RRF - Rouge River Faults							
** AMB - Akron Magnetic Boundary							

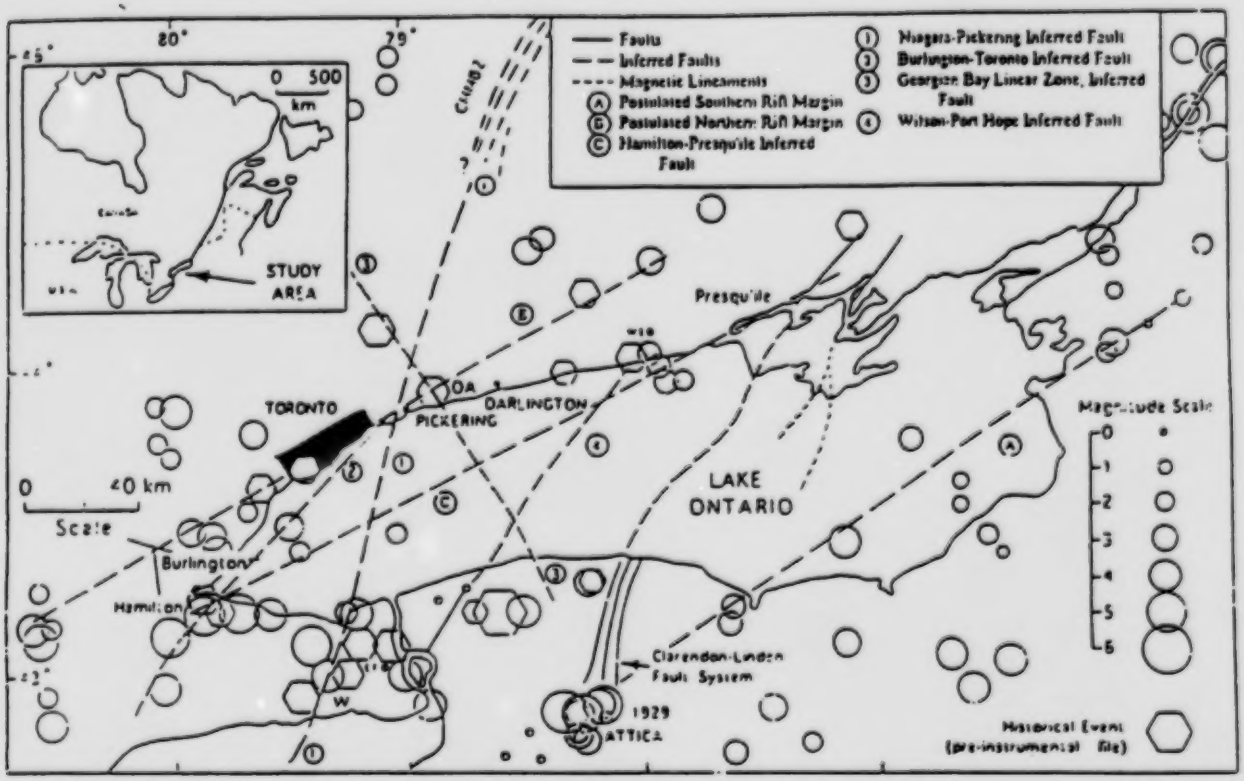


Figure 3-1 Known and postulated structural features of Lake Ontario (from McQuest Marine, 1995)

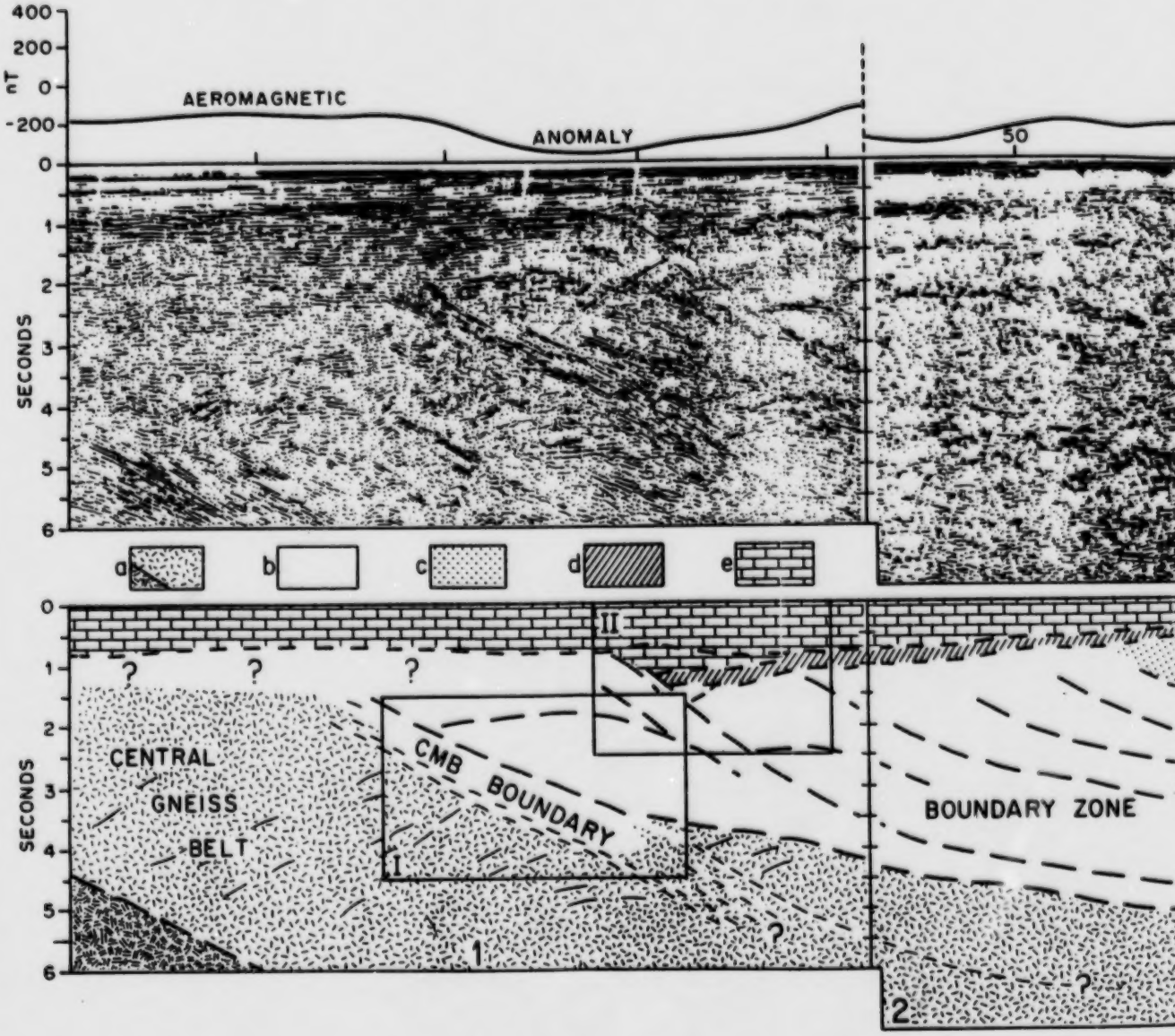


Figure 3-2. East-west composite seismic profile (top) and interpretation (bottom) across the CMBBZ. Note small half-graben in inset box II. *d*, post-Grenville sediments; *e*, Paleozoic strata. Locations of seismic profiles are shown in Figure 2-11. (from Forsyth and others, 1994a)

km
100
0
1
2
3
4
5
6
7
SECONDS
PALEOZOIC SEDIMENTS
CENTRAL
METASEDIMENTARY
BELT
3

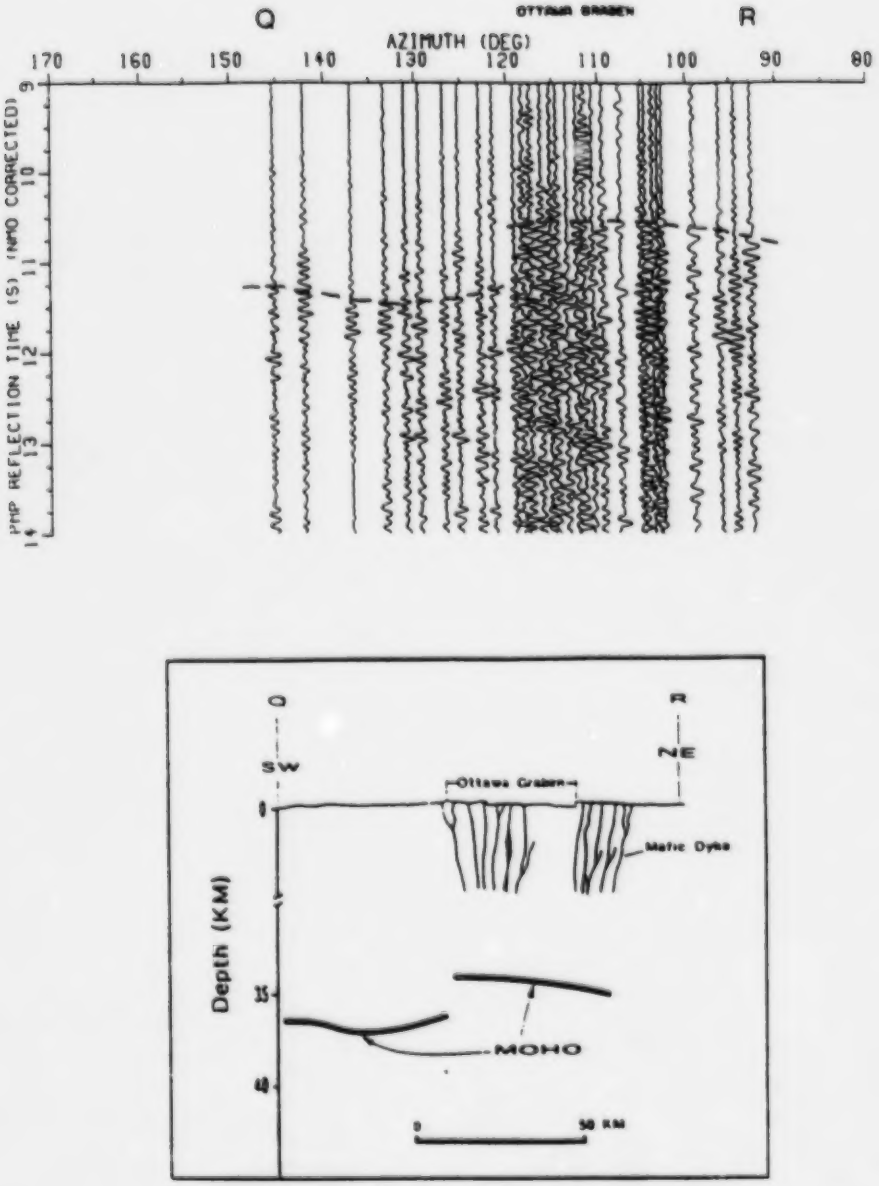


Figure 3-3 Deep seismic reflection data collected across the Ottawa graben (top) and schematic interpretation (bottom). Top figure is from Mereu and others (1986), bottom figure is from Kumarapeli (1993)

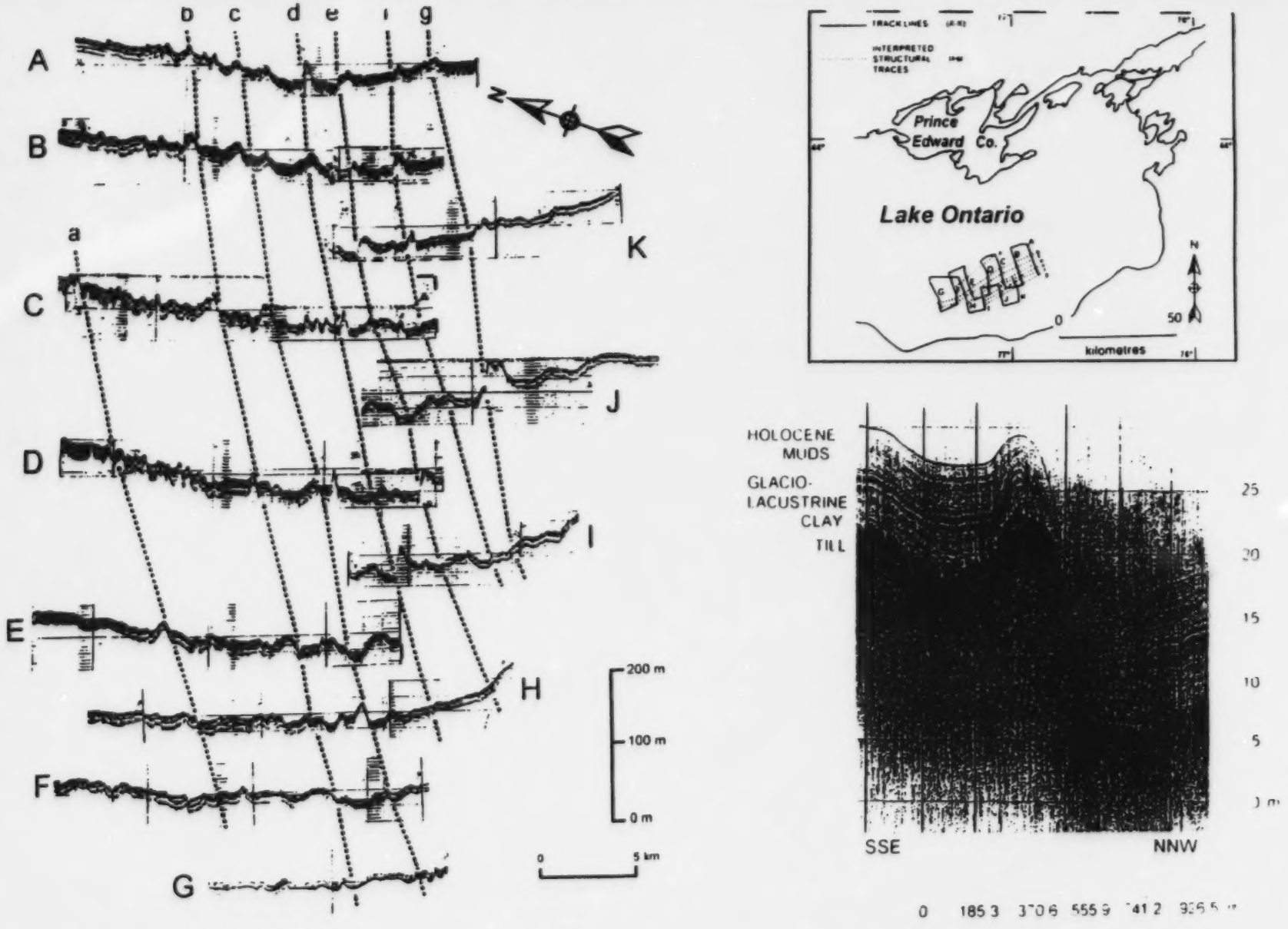


Figure 3-4 Interpreted normal faults from ecograms collected in the South Ontario Structural Zone (left), and a boomer seismic profile across a scarp interpreted as a fault in the SOQZ (lower right. Location of ecogram survey lines and the SOSZ are shown in upper right (from Thomas and other, 1993)

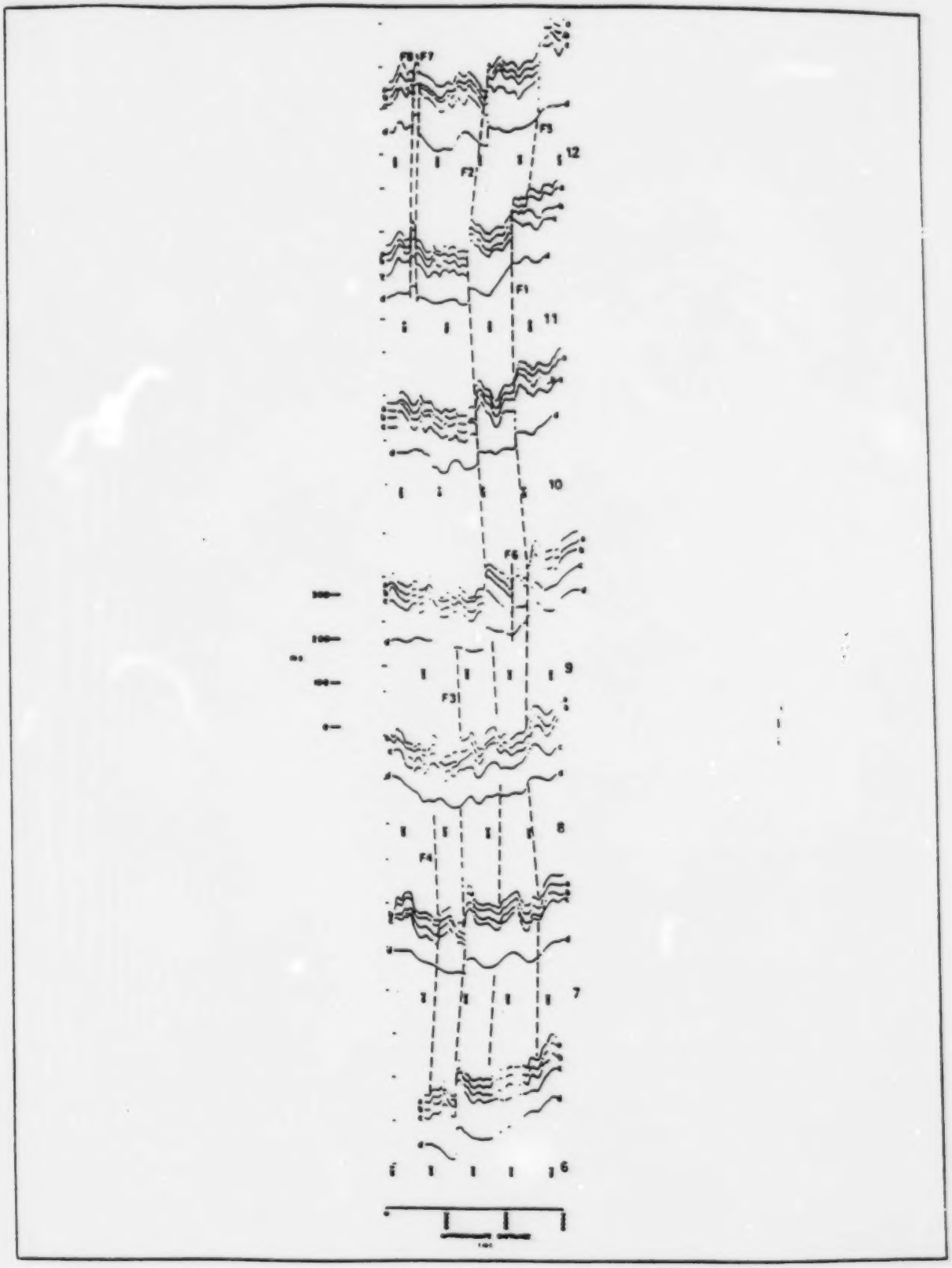


Figure 3-5 Compilation of interpreted seismic lines in the South Ontario Structural Zone (from McQuest Marine, 1995)

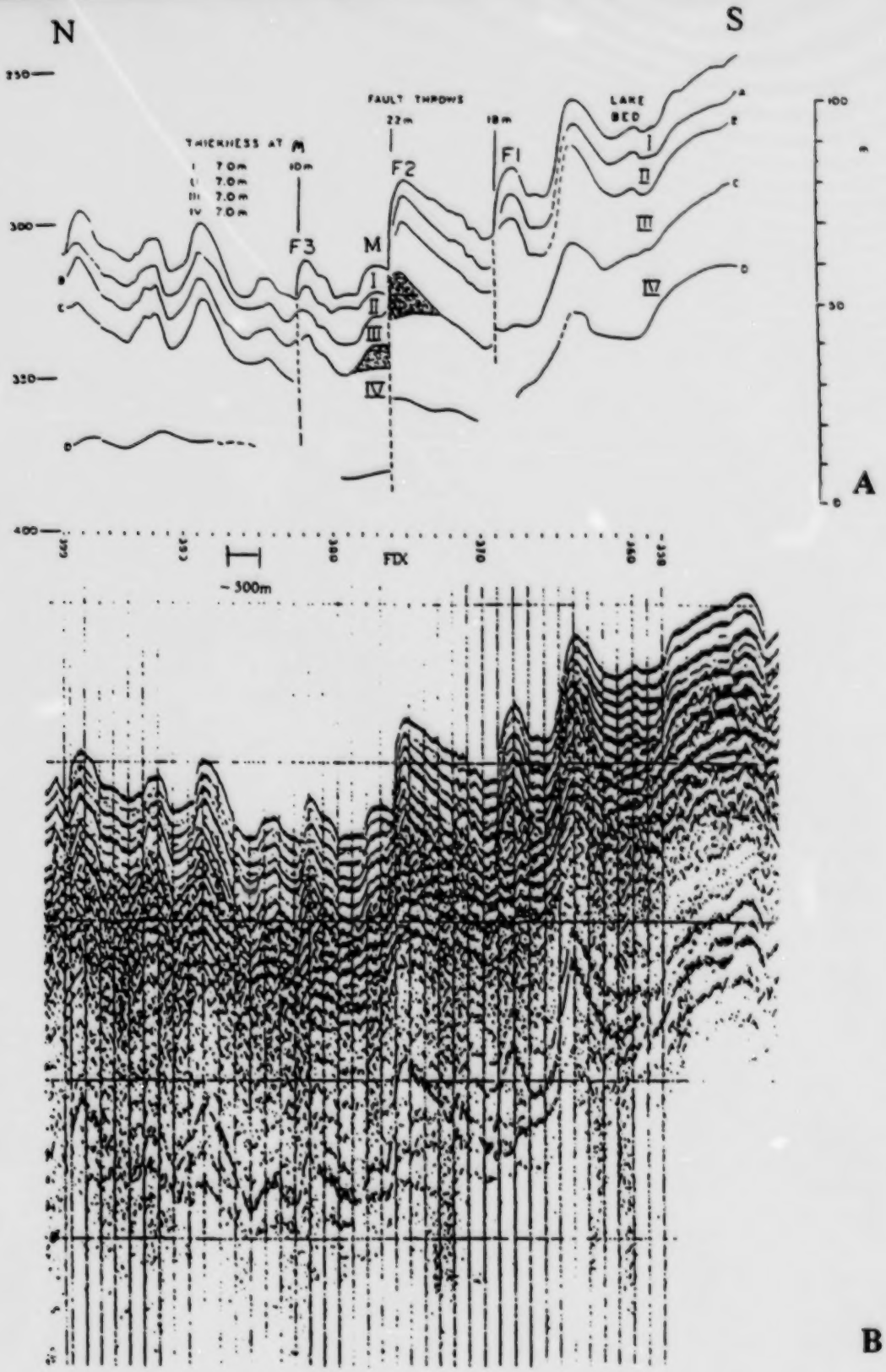


Figure 3-6 Airgun seismic profile (line 9, Figure 3-5) and interpretation (from McQuest Marine, 1995)

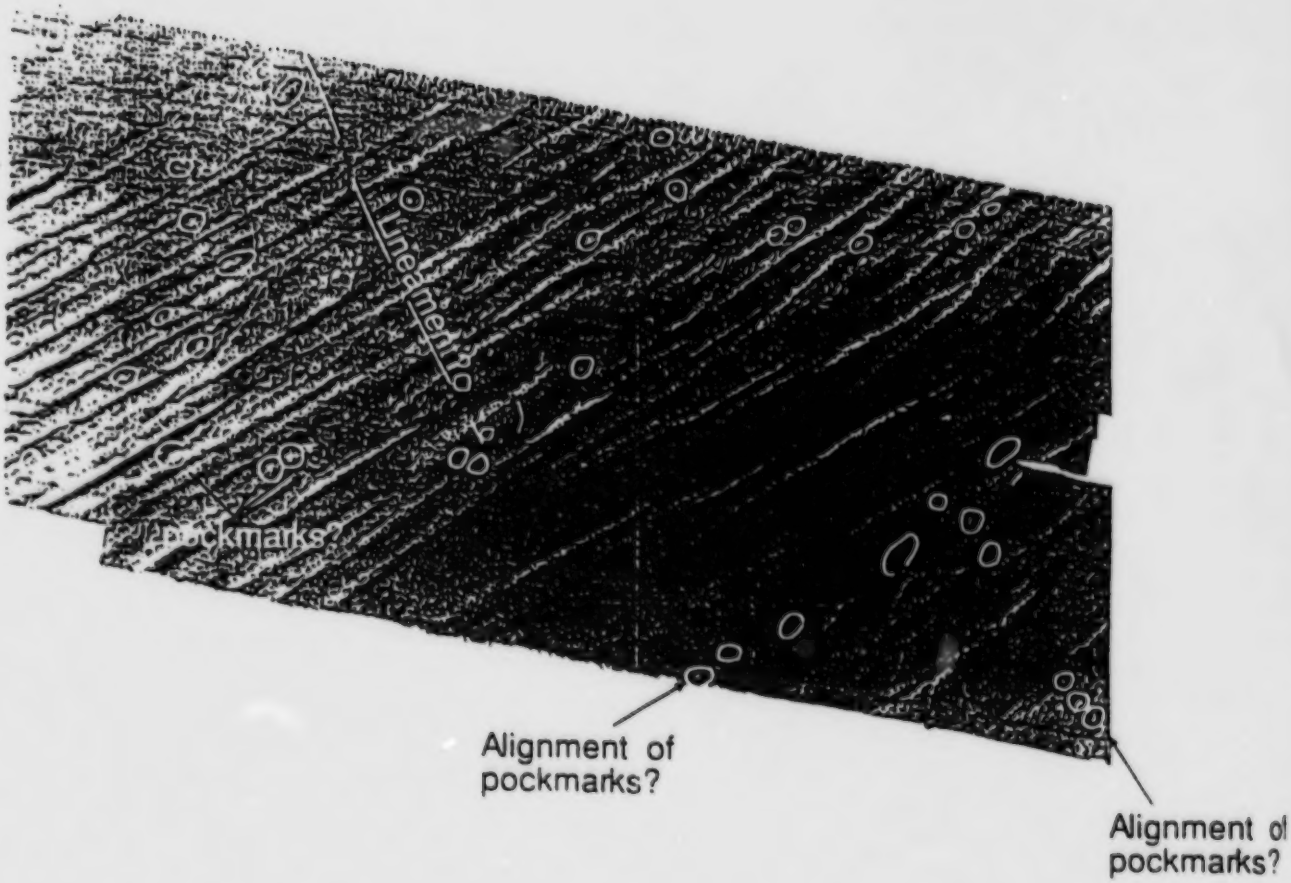
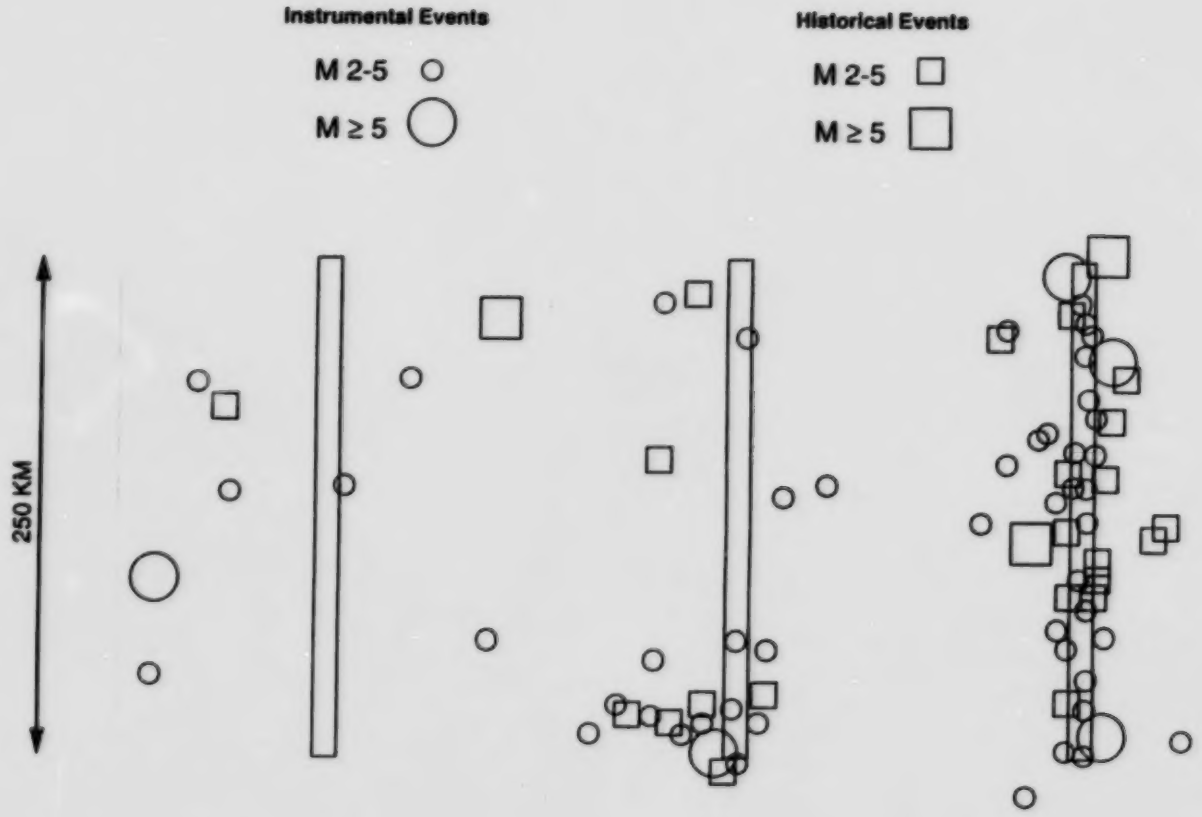


Figure 3-7 Map of swath bathymetry in eastern Lake Ontario showing a prominent field of sub-parallel ridges in a part of the South Ontario Structural Zone. Alignments of pockmarks suggests fluid flow along joints or fractures. (from Lewis, 1995)



	A Assessment	B Assessment	C Assessment
Spatial Association	0.10	0.20	1.00
No Spatial Association	0.90	0.80	0.00

Figure 3-8 Examples of spatial association with seismicity.

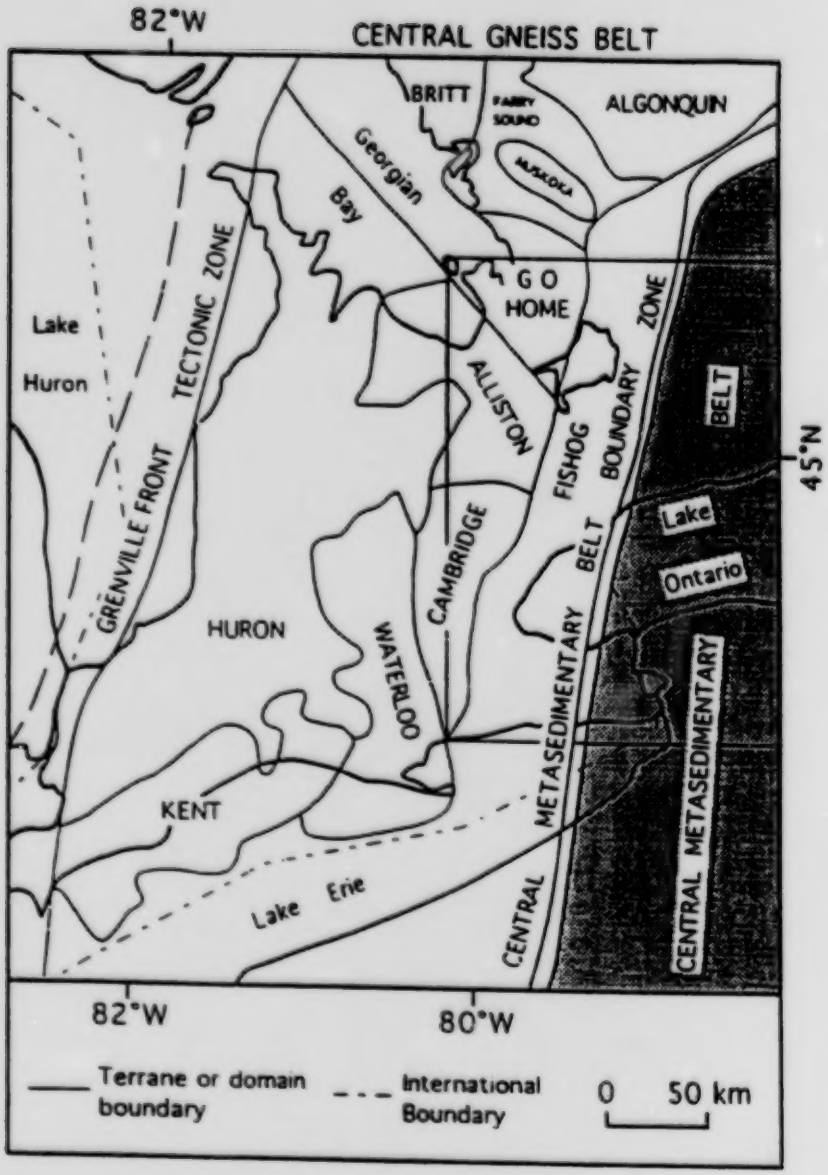


Figure 3-9 Terrane boundaries in the Grenville Province of southeast Ontario (from Easton, 1992).

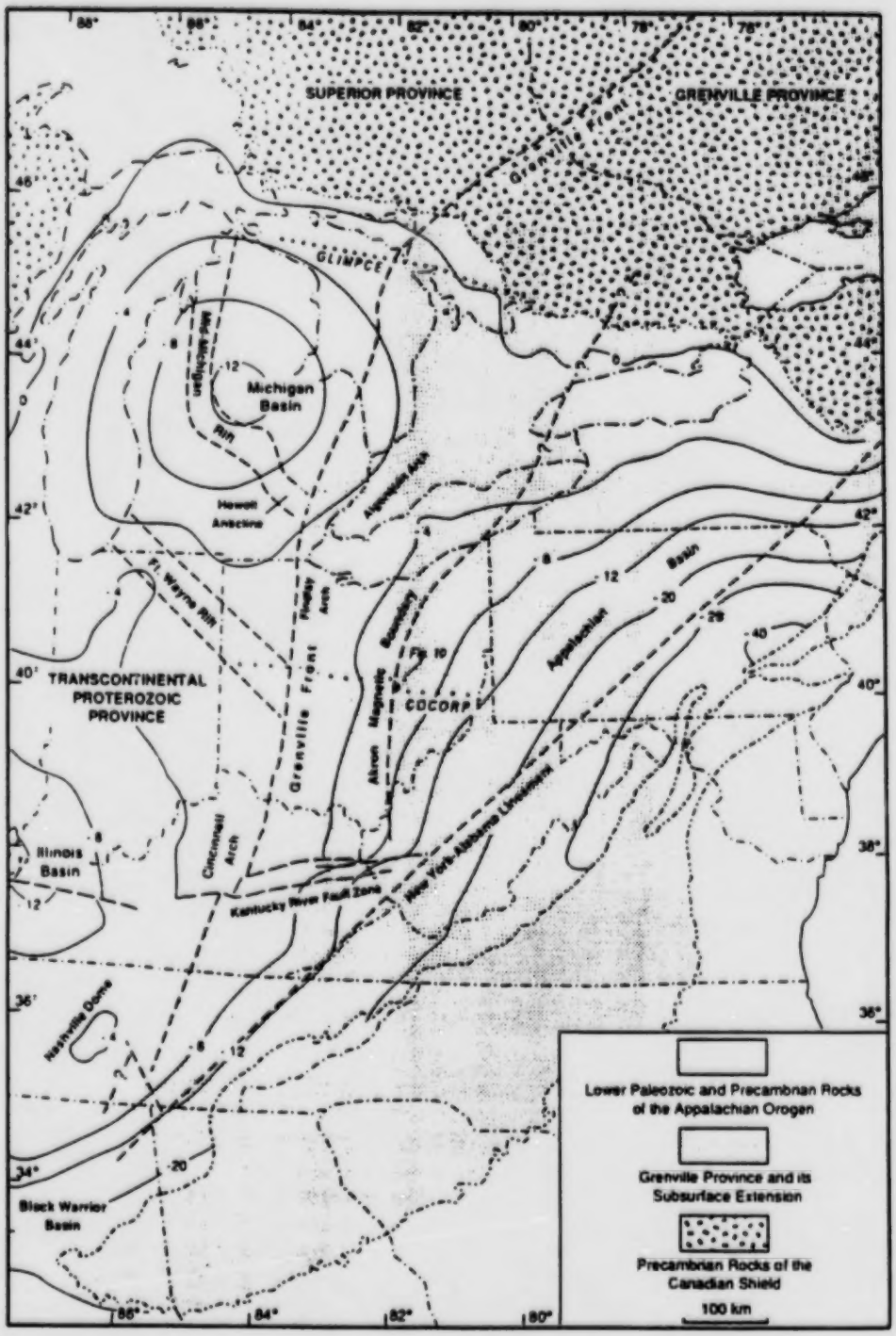


Figure 3-10 Basement configuration map and regional Precambrian and Phanerozoic structures in part of eastern North America. Heavy dashed lines are major basement structures. Note northward continuation of the Akron magnetic boundary appears to coincide with the CMBBZ (from Rankin and others, 1993)

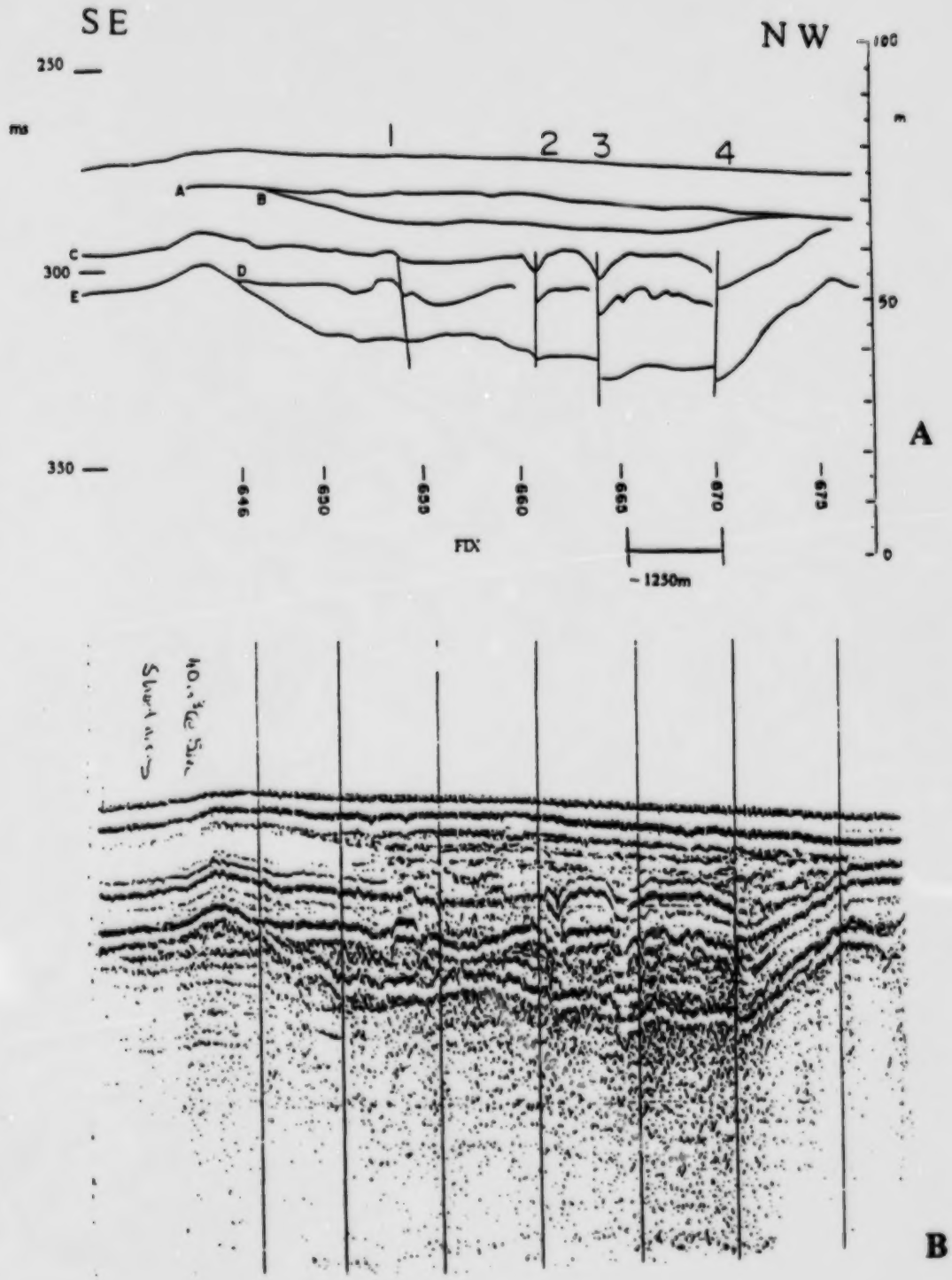


Figure 3-11 Interpreted normal faults in the vicinity of the Niagara-Pickering linear zone and the Hamilton-Presqu'ile fault (from McQuest Marine, 1995)

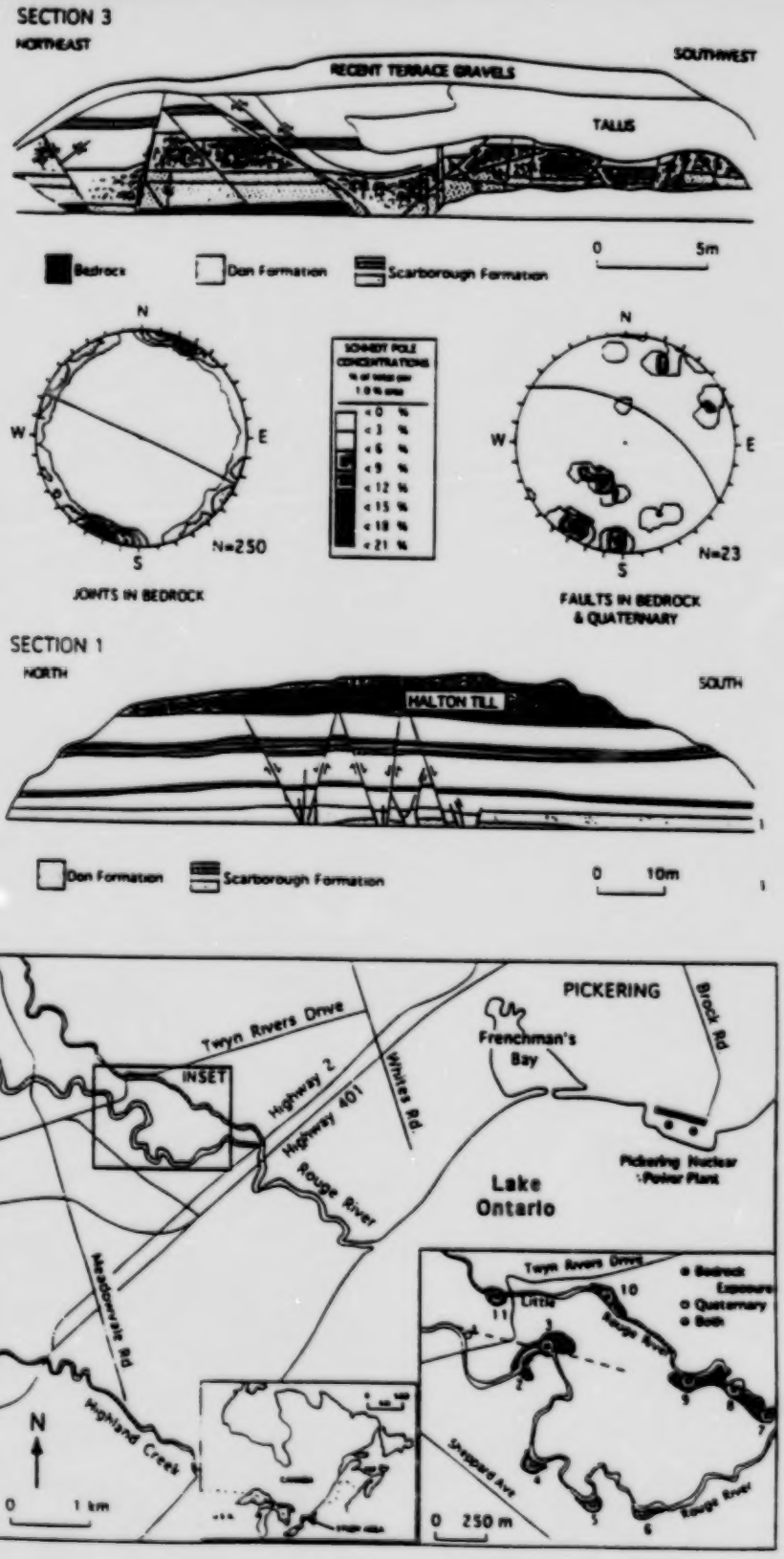


Figure 3-12 Normal faults along the Rouge River (from Mohajer and others, 1992)

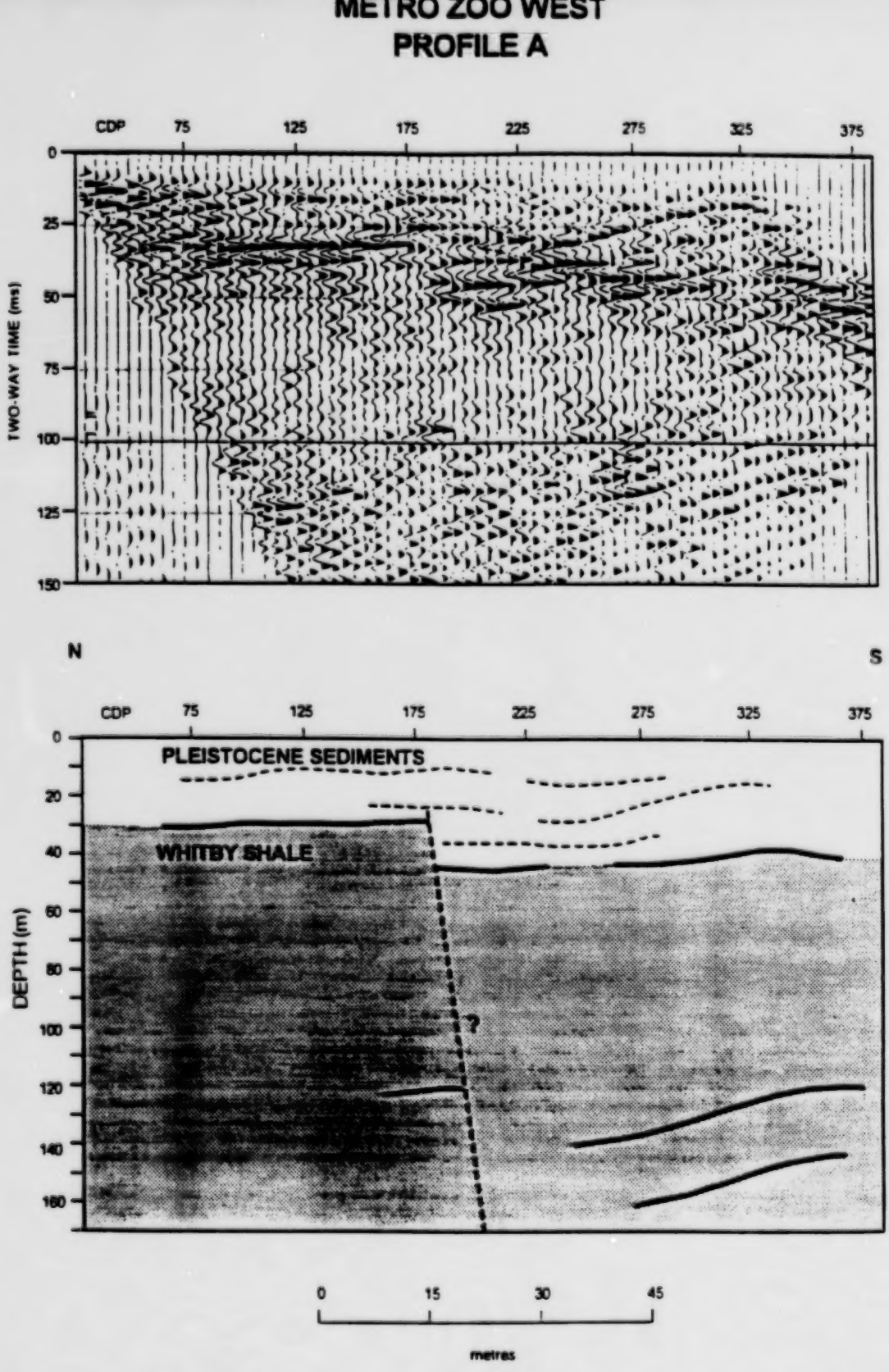


Figure 3-13 Seismic reflection profile in upper Rouge River (top) and interpretation (bottom) (from Mohajer and others. 1995)

4.0 MODELING SEISMIC SOURCES

This chapter summarizes the procedures that are used to model seismic sources as input to the probabilistic seismic hazard anaysis. Because a variety of seismic sources have been identified—representing the range of alternative hypotheses regarding earthquake occurrence in the region—a logic structure was developed that specifies how the various sources should be combined in the hazard analysis. The general framework of the logic structure was discussed in Section 1.1.

In addition to specifying regional and local seismic sources, an alternative approach to specifying the future locations of earthquakes was used: smoothing of observed seismicity. First proposed and utilized as part of the EPRI study (EPRI, 1986) spatial smoothing is now an integral part of the seismic hazard mapping being conducted by the US Geological Survey for the latest round of hazard maps for building code maps of the U.S. under the National Earthquake Hazard Reduction Program. In concept, spatial smoothing is comparable to drawing small source zones around zones of elevated seismicity, but does not require an assumption that the rate of seismicity is uniform within a source zone.

4.1 LOGIC STRUCTURE FOR REGIONAL SOURCES

The logic structure used to define the relationships among regional seismic sources in the site region is given in the logic tree shown in Figure 4-1. (Note: a logic tree expresses the logical sequence of assessments and their dependencies, and alternative models or parameter values are expressed as alternative branches. Relative weights are applied to each branch that express the probability that the particular model or parameter on each branch is correct. Further discussion of logic trees and their use in seismic hazard analyses is given in Coppersmith and Youngs, 1986; 1989). The first assessment is between two alternative approaches to defining the spatial controls on the future locations of seismicity: (1) seismic source zones based primarily on geologic and tectonic bases and (2) approaches and seismic source zones based on observed seismicity. The geologic/tectonically based source zones—although they are derived primarily on the basis of mapped or interpreted major tectonic features (Plates 1 through 10) or crustal lineaments (Plate 11)—are evaluated for their probability of being

seismogenic through a consideration of their association with seismicity. The seismicity-based approaches use the pattern of observed seismicity in a literal sense to define the future pattern of earthquakes. A key difference in these methodologies is the degree to which the spatial pattern of observed seismicity (both historical and instrumentally-recorded earthquakes) provides an indication of the locations of future seismicity. The geologic and tectonic bases for source zonation are strongly favored (0.8) over the seismicity-based alternative (0.2) because it subdivides the region into zones with more uniform crustal characteristics. The need to define sources as areas of uniform seismicity rate is not required because of the techniques that will be used for seismicity modeling within a source can allow for spatially varying rates (see Section 5.2). Thus it is judged that a geologic and/or tectonic basis is more relevant for identifying regions of crust with homogenous characteristics. Of the two geologic approaches, the one based on identification of geologic/tectonic terranes is preferred (0.6) over that based on a composite of lineament patterns and seismicity cluster analyses (0.2) because it defines large crustal areas that have distinct geologic characteristics in terms of their seismogenic potential (e.g. extended vs. nonextend).

The next level of the logic tree addressed the basis for zonation, given a general approach for regional source modeling. For the geologic/tectonic approach, the question addressed is the significance of the boundary between the Iapetan rifted margin (IRM) and the Southern Grenville Region (SGR). The IRM/SGR boundary is judged very likely to be significant to defining regional sources (0.9) because Johnston and others (1994) have shown that the seismogenic potential of a stable continental region varies significantly according to the degree of extension it experienced in the past. For the composite lineament-seismicity approach, only one basis for zonation was used, that employing an analysis of lineament patterns to define similar crustal blocks. Given that seismicity-based approaches are used to describe the future locations of seismicity, there two approaches to zonation are considered. The first is the use of seismic source zones that enclose zones of elevated seismicity. Examples are the Attica seismic zone and the seismicity southwest of Lake Ontario. For this analysis, we use the "seismicity source zones" of Adams and others (1995) (Plate 13). The second approach is spatial smoothing of observed seismicity without imposed source zone boundaries (Plate 12). The methodology used for spatial smoothing of seismicity is one that

smooths both the rate of activity (a-value of the earthquake recurrence relationship) and the b-value. The details of the procedure used are discussed in Section 5 of this report. Advantages of the smoothing approach are that it does not require an interpretation of the boundaries of sources (i.e., the density differences of seismicity are evaluated automatically as part of the procedure) and there is flexibility in the degree to which the seismicity is smoothed. Because of its advantages, the smoothing approach is strongly preferred (0.8) over the seismicity zone approach (0.2).

The third level of the logic tree (Figure 4-1) addresses the location of the IRM/SGR boundary. As discussed in Section 3.1, three alternative locations are considered; one proposed by Wheeler (1995), one proposed for this study (designated WAU on Figure 4-1), and one proposed by Milkereit and others (1992) (designated CMBBZ on Figure 4-1). The weights assigned in Section 3.1 are 0.4, 0.5, and 0.1, respectively. This assessment is not applicable when the IRM/SGR boundary is not considered significant or to the other modeling approaches.

The next assessment in the logic tree (Figure 4-1) is whether or not the St. Lawrence rift system extends southwest of the Ottawa Graben. (Note: the alternative interpretations and weights assigned to each interpretation are discussed in Section 3.1. We here are showing how the various components of the assessment are integrated for purposes of the hazard analysis.) The probability that the St. Lawrence rift extends to the south is judged to be very unlikely (0.1). Assuming, however, that it does extend to the southwest, there are two alternative locations to the northern boundary of the rift zone seismic source zone: along the Hamilton-Presqu'ile fault zone or along a more northerly boundary suggested by McQuest Marine (1995). Because the Hamilton-Presqu'ile fault appears to be a more throughgoing fault zone, higher weight is given to it representing the northern boundary to the St. Lawrence rift extension (0.6) than the more northerly boundary (0.4). The southern extension of the St. Lawrence rift is only considered in the case where the IRM/SGR boundary represents a significant boundary for delineating seismic potential.

The next level of the logic tree addresses the issue of treating the Ottawa and Saguenay Grabens and the St Lawrence extension, if present, as separate sources from the IRM zone (Plates 1a through 9a) or as part of the IRM zone (Plates 1b through 9b). This issue has implications to the assessment of maximum magnitude and earthquake recurrence rates, particularly with regard to the St. Lawrence extension. At this point we are ambivalent as to treating them as separate sources or as part of the IRM zone and assign equal weight to the two approaches.

The final column of the logic tree is not an assessment but is an indication of the various seismic sources that exist given the combination of branches leading to each end point. These are indicated by the appropriate plate number that shows the configurations of seismic sources. For example, follow the path in which: geologic/tectonically-based sources are being used, the IRM/SGR boundary is significant and is located as defined by Wheeler, the St. Lawrence rift does extend to the southwest, the Hamilton-Presqu'ile fault defines the northern boundary of the St. Lawrence rift extension and the St. Larence extension, Ottawa Graben and Saguenay Graben are separate sources from the IRM zones (these are the top set of branches in the logic tree in Figure 4-1). The relative weight associated with this combination of assessments would be the product of the probabilities on all of the branches (i.e., 0.6·0.9·0.4·0.1·0.6·0.5 = 0.0065). The regional seismic source zones that would be present are shown on Plate 4a. (Note that any local seismic sources would be superimposed on these regional source zones, based on their probability of being seismogenic).

The logic tree shown on Figure 4-1 lists only those regional sources that are affected by the alternative interpretations considered in the tree. The hazard analysis will be conducted using all of the sources shown on Plates 1a through 13. For each of these sources, the assessments of seismicity parameters will follow the logic tree format presented in Section 5.3.

4.2 TREATMENT OF ROUGE RIVER FAULT OBSERVATIONS

The observations made in the Rouge River are summarized in Section 3.1. Significant uncertainties exist regarding whether the observed geologic relationships are indicative of tectonic faulting and, if so, which tectonic feature the Rouge River faults might be associated

with. In turn, if the observations are assumed to be associated with a particular tectonic feature, they could have an impact on the assessment of the seismogenic potential of that feature.

The logic structure for considering the Rouge River faults is given in Figure 4-2. The technical basis for the assessments and probabilities on the logic tree are discussed in Section 3.2. The focus here is on the sequencing of the assessments and their implications to sources considered in the hazard analysis. The first assessment is which potential local source should the faults at the Rouge River locality be associated. Three alternative interpretations are considered appropriate to represent the possible seismic source models for the region: (1) the faults might represent a separate seismic source, (2) they might be associated with the Niagara-Pickering linear zone (NPLZ), or (3) they might be part of the Georgian Bay linear zone (GBLZ). For the reasons given in Section 3.2.2, it is judged that the Rouge River faults are most likely a separate feature (0.7), rather than associated with either the NPLZ (0.15) or the GBLZ (0.15).

The next step in the logic tree (indicated by vertical lines) specifies the seismic sources that are considered in the analysis for the particular set of branches that lead to that point. For example, given that the faults are a separate seismic source, the sources considered are the Rouge River faults, the NPLZ, and the GBLZ. Alternatively, if the Rouge River faults are associated with either the NPLZ or the GBLZ, then these are the only two sources.

The next assessment is whether or not the Akron magnetic boundary should be considered part of the Niagara-Pickering linear zone. The arguments for and against this association are presented in Section 3.2 in the discussion of the seismogenic potential of the Niagara-Pickering linear zone.

Finally, the assessment of the probability of being seismogenic is made conditional on the association of the Rouge River faults with the other two local sources and on the association of the Akron magnetic boundary with the NPLZ. The assessments shown on the logic tree (Figure 4-2) are summarized in Table 3-3 and discussed in Section 3.2.

The logic tree shown on Figure 4-2 lists only those local sources that are affected by the alternative interpretations considered in the tree. The hazard analysis will be conducted using all of the local sources shown on Plates 14 and 15. For each of these sources, the assessments of seismicity parameters will follow the logic tree format presented in Section 5.3.

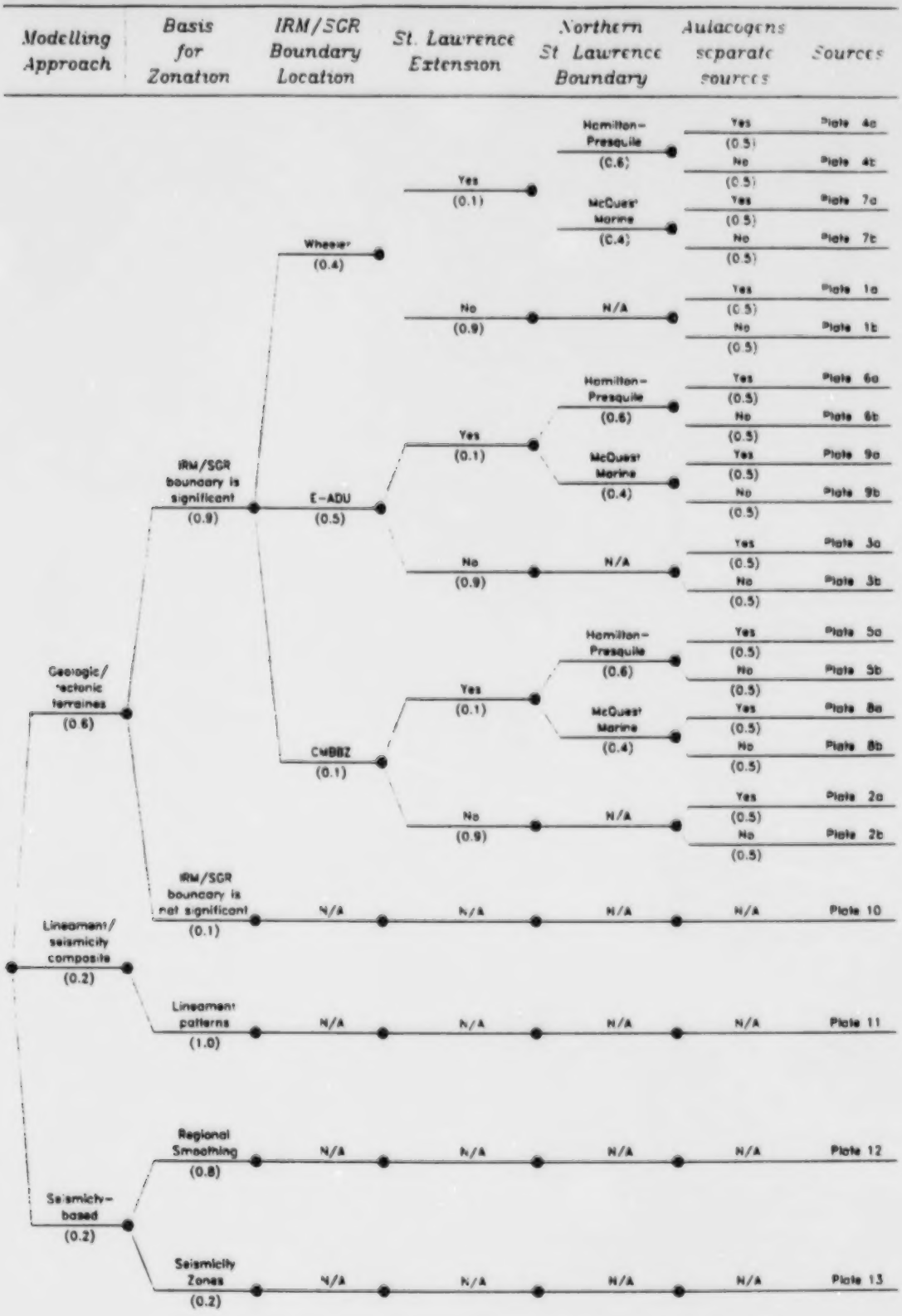


Figure 4–1 Logic tree showing relationships among regional sources. Given the branches that lead to the end of the tree, the "source" column identifies the plate showing the sources that will be used in the PSHA.

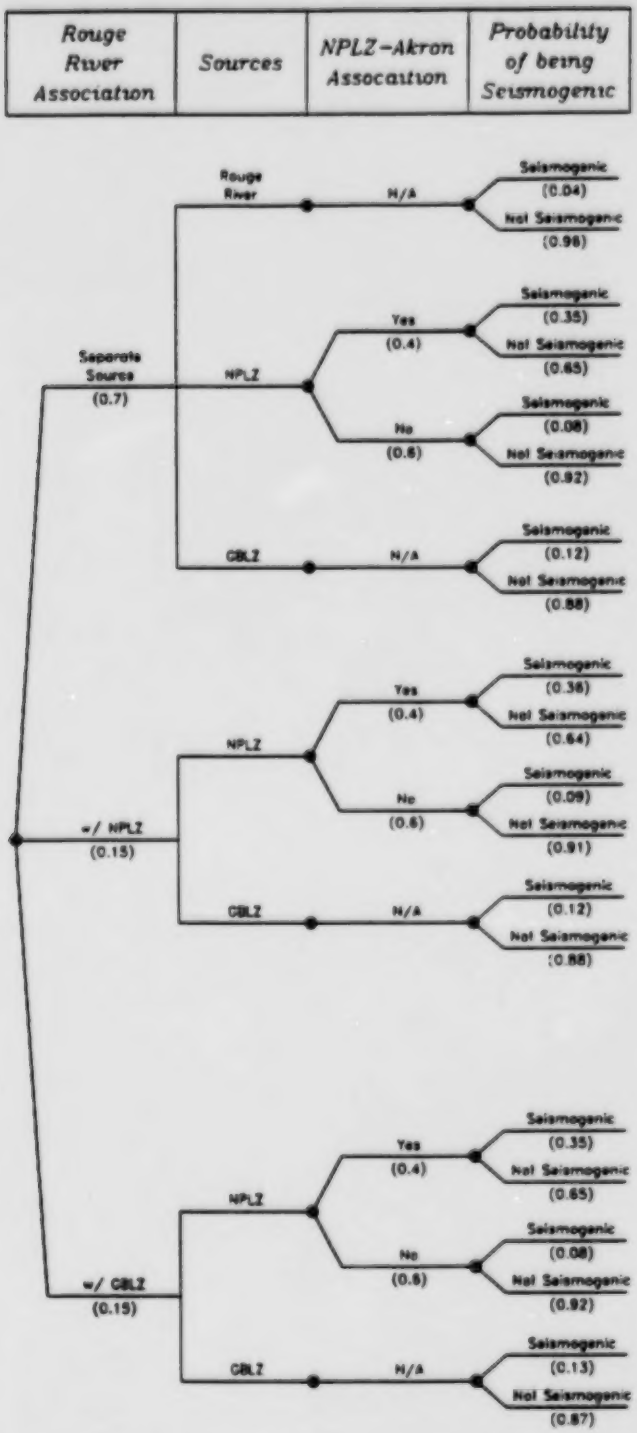


Figure 4–2 Logic tree for Rouge River, Niagara–Pickering linear zone (NPLZ) and Georgian Bay linear zone (GBLZ) local sources showing their potential associations and the probabilities of each source being seismogenic.

5.0 APPROACHES TO ASSESSING MAXIMUM MAGNITUDES AND EARTHQUAKE RECURRENCE

Several types of seismic sources are considered in this study. The regional source zones shown on Plates 1 through 11 are defined on the basis of the geology and tectonics of the region. These sources represent regions of the earth's crust that are believed to have homogeneous characteristics with regard to earthquake occurrence. The regional seismicity zones shown on Plate 13 represent areas in which the observed seismicity patterns appear to be uniform. As indicated on Figure 4-1, the seismicity zones are considered to be an alternative to the geologically-based regional source zones. Superimposed on the regional sources or seismicity zones are the local sources. The local sources may either represent local concentrations of seismicity caused by (or indicated by) some particular tectonic feature or they may represent actual throughgoing fault zones. This different interpretation of local source zones lead to differences in the characterization of maximum magnitude and seismicity parameters.

Discussed in this chapter are the methods that are planned to be used for assessing maximum magnitudes and earthquake recurrence parameters for these sources. The intent is to identify the appropriate approaches. The first two subsections present the approaches for maximum magnitude and seismicity parameters. The third subsection outlines how these methods will be applied to the various types of seismic sources.

5.1 Maximum Magnitude Approaches

Maximum earthquake magnitudes (i.e., the upper-bound magnitude in the frequency-magnitude relationship) must be estimated for each seismic source. Experience in seismic hazard assessments in eastern North America have shown that the estimation of maximum magnitudes is usually highly uncertain. Therefore, the maximum magnitude estimate for a particular seismic source is expressed as a probability distribution that reflects this uncertainty.

In more active seismic environments, such as the western Unites States, maximum magnitudes are commonly estimated based on geologic evaluations of the dimensions of fault ruptures.

such as maximum fault rupture length, rupture area, displacement per event, etc. These dimensions are then correlated with earthquake dimensions based on empirical regressions (e.g., Coppersmith, 1991; Wells and Coppersmith, 1994). Despite the relative stability in the *methods* used to assess maximum magnitudes in these tectonic environments, there is still significant uncertainty in the assessments of rupture dimensions for a given fault and in alternative methods used to make the assessment. As a result it is common for the maximum magnitude probability distribution for a particular fault source in a PSHA to span 0.5 to 1 magnitude units. It should be expected, therefore, that the uncertainties in maximum magnitudes for seismic sources in the East should be still greater than in the more active environments, because the causative faults are usually not known.

Two principal approaches are used to estimate maximum magnitude for seismic sources in this study: the first is the approach developed as part of the EPRI Stable Continental Regions study (Johnston and others, 1994), and the second, which is applied only to fault-related seismic sources, is based on estimated rupture lengths. These approaches and their applicability to seismic sources in this study is summarized below.

5.1.1 EPRI Maximum Earthquake Approach

A recently completed analysis of large earthquakes in stable continental regions (SCR) sponsored by the Electric Power Research Institute (Johnston and others, 1994) specifically addressed the problem of maximum magnitudes. The SCR earthquake project developed worldwide data bases that could be used to make scientifically-supportable assessments of maximum earthquake magnitude for seismic sources in eastern North America for use in probabilistic seismic hazard analysis. The approach developed in the study is based on the notion that the maximum magnitude that a seismic source is capable of generating is rare relative to the short historical record and it is unlikely that the historical record includes the maximum event. Therefore, space is substituted for time by looking globally at SCR regions and documenting the size of earthquakes that have occurred within tectonic environments similar to those in eastern North America.

Johnston and others (1994) developed a consistent set of observations about the tectonics of stable continental regions worldwide and the nature of large earthquakes that occur within them. They recommend a Bayesian approach for assessing maximum magnitude that is based on a prior distribution of maximum magnitude coming from the statistical analysis of the global data base, updated with information (the sample likelihood function for maximum magnitude) that is related specifically to the seismic source of interest. The final product is a probabilistic distribution of maximum magnitude that incorporates uncertainties of the assessment. This probabilistic distribution obtained using the Johnston and others (1994) approach is in terms of the moment magnitude, M. To be consistent with the magnitude m_b adopted in this study, we converted the M-based distribution to an m_b-based distribution using the magnitude conversion relation given in Atkinson and Boore (1995).

The information needed to estimate the maximum magnitude distribution in a source zone using the Johnston and others (1994) technique include: (1) the list of descriptor values; (2) the observed, historical maximum magnitude; and (3) the number of historical earthquakes above moment magnitude 4.5. The first item is needed for setting up the prior distribution, whereas the last two items are source-specific information needed to compute the sample likelihood function of maximum magnitude. Detailed formulations are given in Sections 5 and 6 of Johnston and others (1994).

The prior distribution is assumed to be a Gaussian distribution whose mean and standard error are derived from the statistical analysis of the global data base. Johnston and others (1994) provide two approaches to developing the prior distribution. One is based on multivariate regression analysis that considers the influence of a number of variables on the mean of the distribution. Table 5-1 describes these variables and Table 5-2 gives the coefficients and standard error of the multivariate regression model. Johnston and others (1994) note that the results of this analysis suggest that none of the descriptors are a strong predictor or determinant of maximum magnitude (low t-statistic), although geologic age and the existence of a compressional stress state show some mildly significant effects (Table 5-2). They conclude that the net benefit of using the multivariate model is low as suggested by the coefficient of determination, r^2, given in Table 5-2.

The second approach to developing a prior distribution considers only the crustal type. The data base is grouped into extended and non-extended types. The bias-adjusted estimates for the mean maximum magnitude are M 6.3 and 6.4 for non-extended and extended crust, respectively (Johnston and others, 1994). The standard errors estimated for non-extended and extended crust are 0.5 and 0.84 (Johnston and others, 1994).

In this study, we develop prior distributions using both approaches, each approach is given the same weight (0.5). The final distribution of maximum magnitude for each of these source zones is obtained by multiplying the prior with the sample likelihood function based on the observed maximum magnitude and the number of events greater than and equal to **M** 4.5 within the source zone. Figures 5-1 and 5-2 illustrate the process. Shown at the top of the figures are the prior distributions for maximum magnitude for non-extended and extended crust. The middle plots show sample likelihood functions for maximum magnitude given a maximum observed event of **M** 5.5 and either one event larger than **M** 4.5 (Figure 5-1) or 10 events larger than **M** 4.5 (Figure 5-2). The bottom plots show the resulting posterior distributions for maximum magnitude. We will then discretize the posterior distributions at intervals of 0.25 magnitude units to obtain a discrete distribution centered on one-half magnitude values.

Figure 5-3, part a, shows an example of a maximum magnitude distribution for a hypothetical source zones in non-extended crust in which three earthquakes of moment magnitude greater than 4.5 have been recorded, the largest having a moment magnitude of 5.5. The maximum magnitude distribution developed from applying the EPRI approach was converted to m_b using the relationship given in Atkinson and Boore (1995).

The advantages of the EPRI approach are that it is based on a global analysis of SCR earthquakes and it takes into account seismic-source specific information (i.e., the largest earthquake that has been observed within each seismic source and the number of **M**>4.5 earthquakes). The resulting probabilistic maximum magnitude distribution provides a realistic assessment of uncertainty.

5.1.2 Fault Rupture Dimensions

As noted previously, assessing the dimensions of fault rupture is difficult, especially in tectonic environments such as eastern North America. This is because the location and extent of reactivated/active faults is usually not known. Instead, regional faults and other structures (associated with previous tectonic regimes) are known or interpreted from various data sets. The lengths and geometries of these regional structures can provide some indication of possible maximum rupture dimensions. Significant uncertainties are related to the manner in which the long regional structures might actually be reactivated in the present regime and the segmentation that might exist.

For this study the estimates of possible rupture lengths given in Mohajer (1995b) for various tectonic features will be used as one approach for the local seismic sources when they are considered to represent faults. Mohajer (1995b) provided estimates of rupture length based on a consideration of segmentation of the regional fault features, using primarily geometric segmentation criteria (e.g., see discussion of segmentation in Coppersmith, 1991). He provided mean estimates of earthquake magnitude for each segment using the Wells and Coppersmith (1994) empirical rupture length regression. To account for uncertainty in the assessment of maximum rupture dimensions and selection of the appropriate empirical attenuation relationship, we will use these mean estimates and assume a distribution ranging over plus or minus 0.5 magnitude units . Figure 5-3, part (b), shows an example of a maximum magnitude distribution for a hypothetical local source where the estimated maximum rupture length is 40 km. From Wells and Coppersmith's (1994) empirical relationship between rupture length and magnitude, the expected magnitude for a 35-km rupture length is moment magnitude M 6.7. This is approximately equal to m_b 6.75. Using this as the midpoint of the maximum magnitude distribution, a relatively broad discrete distribution shown on the figure is defined such that the standard error in maximum magnitude is approximately 0.25 magnitude units.

In the case of the Rouge River fault, if it is assumed to be a tectonic fault, the estimates of cumulative displacement in the exposure is about 1.25 to 4 m. It is also interpreted that multiple displacements are present in the exposures. Therefore, a displacement per event of

about 1 m is estimated for the fault and this value is used with empirical relationships between displacement per event and magnitude to estimate maximum magnitude.

5.2 EARTHQUAKE RECURRENCE APPROACHES

5.2.1 Earthquake Catalog Analysis

The principal database for assessing earthquake recurrence is the historical and instrumental seismicity record. The primary source for the earthquake catalog used in this study is the seismicity catalog provided by the Geological Survey of Canada for the period from 1500 through June of 1996. This catalog will be supplemented with the NCEER91 catalog (Seeber and Armbruster, 1991) covering the period from 1627 to 1985 and data from the U.S. National Earthquake Information Center (NEIC) for the time period post 1984. The primary magnitude measure that will be used is the Lg body wave magnitude, m_{bLg}, which is considered to be equivalent to Nuttli magnitude, m_N and to body wave magnitude m_b. The designations of m_b in the two catalog sources are very consistent. Figure 5-4 shows a comparison of the m_b designations for 1,451 common events in the two catalogs. The best fit regression line has a slope of 1, and intercept of near zero, and the 90-percent confidence interval on the regression line is indistinguishable from a 1:1 line. The remaining magnitude assessments in the GSC catalog will be converted to m_{bLg} using the approach outlined in EPRI (1986).

Differences in the earthquake locations listed in the GSC and NCEER91 catalogs were examined for earthquakes in the site vicinity. Figure 5-5 shows the earthquake locations given in the GSC and NCEER91 catalogs with lines connecting the two locations for the same event. There are significant differences in the listed locations for only six events within the site vicinity. The dates for these events are indicated on Figure 5-5. For all of the remaining events listed in both catalogs, the locations are either identical or within the location accuracy. Of the six events with major differences in location, only the difference for the location of the 1857 Lockport earthquake is significant to the assessment of the seismic potential of the local sources. This event is discussed in Section 3.2.2.

The formulation used in this PSHA assumes that the occurrence of earthquakes conforms to a Poisson process, implying independence between the time of occurrence of events. Thus, it is necessary to remove dependent events from the counts of earthquakes used for recurrence evaluations. We intend to use the local cluster analysis procedure developed by EPRI (1986). In this process, the spatial-temporal extent of a cluster is first identified; then, the largest earthquake in the cluster is classified as main earthquake and the remaining members of the cluster as dependent events. The cluster region is estimated as the largest spatial-temporal window (surrounding the main event) in which the seismicity rate is statistically different from the background rate. The shape and size of a cluster thus varies from one cluster to another. This approach is more flexible than the traditional approach where the extent of a foreshock/aftershock cluster depends only on the magnitude of the main earthquake.

The earthquake sequence in a historical catalog is normally nonstationary in time due to incomplete reporting of earthquakes. The standard approach of dealing with incomplete reporting is to estimate the recurrence rate based on the seismic activity during a time period over which earthquakes can be considered completely reported in the catalog (e.g. Weichert, 1980). This approach requires that earthquakes occurred in the period of incomplete reporting not be used. The completeness period is usually determined by a method based on the stability of observed recurrence rate in a magnitude interval as a function of time before the present (Stepp, 1972). The point in time at which the rate begins to steadily diminish with increasing time before the present indicates the time before which the catalog may be incomplete. This method requires a certain degree of judgment at very low and very high magnitudes.

For this study, we intend to use the general recurrence model developed by EPRI (1986) which allows incompletely reported data to be used in the estimation of recurrence rate. The basic component of the EPRI (1986) model is the use of probability of earthquake detection to account for the reduction of observed rate due to incomplete reporting. The rate of the "catalog" earthquake is the product of the true rate and the probability of detection, which is defined as the probability that an earthquake of given magnitude m_k, origin time t, and epicenter x is reported in the catalog. We intend to repeat the analysis presented in EPRI

(1986) using the updated catalog of earthquakes developed for this study. We will compare these results with those obtained using the more traditional approach defined by Stepp (1972).

5.2.2 Earthquake Parameter Assessment for Regional Source Zones

Recurrence relationships will be developed for all regional seismic source zones by incorporating the events that lie within the source boundaries. We assume that the standard truncated exponential distribution applies to the occurrence of independent earthquakes between the minimum magnitude of interest and the maximum magnitude for the source zone. Furthermore, we assume that the earthquakes occurring within a regional source zone represent a homogeneous population that can be represented by a single magnitude-frequency distribution (recurrence curve). The seismicity parameters of the recurrence relationship for a source zone will be estimated using the maximum likelihood formulation presented by Weichert (1980) and the number of recorded events in various magnitude intervals within the source zone. The maximum likelihood formulation provides not only a best estimate of the recurrence parameters, but also an assessment of their uncertainty. Figure 5-6 illustrates the approach that will be used to incorporate the uncertainty in the parameters into the PSHA. For each source zone and maximum magnitude assessment, 25 sets of recurrence parameters will be developed (5 alternative values for seismicity rate and 5 alternative *b*-values) using the uncertainty estimates derived from the maximum likelihood assessment for the zone. The relative likelihood for each of these parameter sets will be computed using the likelihood formulation. The relative weights will be used to define a discrete probability distribution for the seismicity parameters. The resulting recurrence rate distribution (including the uncertainty in maximum magnitude) is illustrated on part (b) of Figure 5-6. This approach assures the proper probabilistic dependency between the seismicity rate parameter and the *b*-value warranted by the data.

Most past PSHA studies have assumed that the spatial distribution of earthquakes is uniform within a source zone. We propose to relax that assumption for this study by considering an alterative approach in which the spatial density of earthquakes within a source zone is defined using kernel density estimation. This approach is currently being used to develop the national seismic hazard maps for the U.S. (Frankel, 1995). All of the recorded events, regardless of

size, are used to form a spatial density function using standard kernel density estimation techniques. Figure 5-7 (taken from Chapman and others, 1996 in press) illustrates a spatial density estimate using a kernel density approach for a source region in eastern Tennessee. The important parameter in this process is the characteristic dimension, h, of the smoothing kernel. We will estimate values for h from the data using techniques presented in Silverman (1978). Uncertainty in h will be incorporated into the analysis.

EPRI (1986) also developed an approach for modeling spatially varying seismicity parameters. In their approach the region of interest (a source zone or the entire region) is subdivided into individual cells with a dimension of one or one-half degree in longitude or latitude. The seismicity parameters for each subzone are estimated from the earthquake event counts within each cell using the maximum likelihood method. The seismicity parameter estimates are subject to an imposed degree of smooth variation in the parameters across the cell boundaries. The smoothing is implemented by imposing a penalty for deviation of the parameters of an individual cell from the average for the surrounding cells. The larger the penalty, the greater the degree of uniformity of parameters among the individual cells. In the limit, a very high penalty results in uniform seismicity parameters. Figure 5-8 (taken from Geomatrix, 1995) illustrates the application of this process to a regional source zone along the eastern coast of the U.S. The effect of varying the degree of smoothing is illustrated by the results for different penalty values. The EPRI (1986) methodology provides a means of assessing the uncertainty in the recurrence parameters by simulating possible parameter sets using the statistics of the fit to the observed data. Figure 5-9 illustrates an example of such a simulation for several of the subzones of the regional zone shown on Figure 5-8. The EPRI (1986) methodology for spatially varying seismicity parameters implies that the earthquakes within the region analyzed do not necesarily belong to a homogeneous population defined by a single magnitude frequency distribution. Thus it is particularly suited to cases when no source zone boundaries are imposed.

5.2.3 Earthquake Recurrence Parameters for Local Sources

Recurrence analysis for the local sources takes into account uncertainties in both the location of the faults and in epicentral locations. Because the local sources (e.g., the NPLZ, GBLZ) are

identified on the basis of geophysical data and not on continuous exposures along strike of faulting and displacement, the location of the actual seismicity concentrator or fault is uncertain within about a 10 km zone. Therefore the local sources (with the exception of the Rouge River fault) are modeled as occurring within a 10km-wide zone for the hazard analysis. For calculating recurrence relationships, seismicity is gathered within a corridor that extends 25 km either side of the source (resulting in a 60 km wide zone). For purposes of the hazard analysis these events are assumed to have occurred on the local source.

As discussed in Section 4, the Rouge River fault is modeled alternatively as being a separate seismic source, a part of the NPLZ, or part of the GBLZ. At present, it is not known what the extent of the Rouge River fault is (there are two locations that are about 500 m apart). If the Rouge River fault is tectonic and is a separate seismic source, its length is estimated based on the geologic observations in the exposure. The available data suggest a displacement per event of about one meter (if the displacements are tectonic). Using empirical relationships between displacement and rupture length, this would imply a length of about 30 km. The orientation of this assumed 30 km source is estimated based on the observed fault orientations measured at the exposure. In addition to the use of seismicity data, earthquake recurrence rate is also estimated for the Rouge River fault based on its assumed slip rate. Estimates of the displacement observed in the exposures of the fault range from about 1.25 to 4 m and the estimates of the age of the deposits range from 40,000 to 12,000 yr. These displacements and ages imply a range of slip rates from 0.3 to 0.03 mm/yr. This slip rate is used with the estimates of its length to develop a recurrence relationship for the source, assuming it is seismogenic.

If the local source zones are considered to represent throughgoing faults, then it is uncertain whether or not the standard truncated exponential recurrence relationship applies to them. Studies of individual faults in more active tectonic regions have indicated that an alternative recurrence model, the "characteristic" magnitude distribution, is more applicable to individual faults and fault segments. Youngs and Coppersmith (1985) presented a magnitude density relationship for the characteristic magnitude recurrence model that specifies the relative frequency of small, exponentially distributed events and large, characteristic events that are

near the maximum magnitude in size. We propose to use both the characteristic and truncated exponential recurrence models to compute seismicity rates for the local sources in the PSHA. Figure 5-10 compares truncated exponential and characteristic magnitude distributions (as defined by Youngs and Coppersmith, 1985) fit to observed small magnitude seismicity.

5.3 Seismicity Parameter Charaterization for Seismic Sources

Figures 5-11 through 5-14 illustrate the logic trees that will be used to characterize the seismicity parameter assessments for the four types of seismic sources.

Figure 5-11 shows the logic tree for regional source zones (Plates 1a through 10). The source model for the regional sources is that of a zone rather than a fault. The EPRI (1994) methodology will be used to estimate maximum magnitude, resulting in a distribution for maximum magnitudes such as the one shown on part (a) of Figure 5-3. The approach illustrated on Figure 5-6 will be used to define the distributions of seismicity parameters for the zone and a truncated exponential magnitude distribution will be used. Two alternative spatial densities will be used, uniform and a kernel density estimate. The kernel density estimate is strongly preferred (0.9) to the uniform model (0.1) because most of the zones exhibit evidence of clustering of the seismicity.

Figure 5-12 shows the logic tree that will be used to characterize the seismicity parameter assessments for the local sources (Plates 14 and 15). The first assessment is whether to consider the local source to be a concentrator of seismicity, and thus have the characteristics of a zone, or to be a throughgoing fault. The weight assigned to the concentrator (zone) model versus the throughgoing fault model is dependent on the local source being considered. The Georgian Bay Linear Zone and the Toronto-Hamilton seismic zone appear to have no single fault-like feature and the zone model is strongly favored (0.95) over that of the throughgoing fault model (0.05). At the other end of the spectrum, the Clarendon-Linden fault system and the Rouge River faults have been identified as fault systems, and the throughgoing fault model is favored (0.8) for these local sources over the zone model (0.2). The Niagara-Pickering Linear Zone, the Hamilton-Presqu'ile fault and the Wilson-Port Hope Lineament do not display clear evidence of behavior as throughgoing fault in the present tectonic regime and the

zone model is favored (0.8) over the fault model (0.2) for these sources. If the local source is considered to be a zone, then the EPRI (1994) approach for maximum magnitude applied to the regional zone in which the local source lies will be used to define the maximum magnitude distribution. The seismicity parameters will be determined using the approach illustrated on Figure 5-6 and the truncated exponential magnitude distribution will be used. If the local source is considered to be a throughgoing fault, then rupture dimensions will be used to assess the maximum magnitude. The seismicity rate parameters will be assessed in the same way as when the source is considered to be a zone, except that the characteristic magnitude distribution is assumed to apply. The seismicity data associated with any single local source is unlikely to provide a tight constraint on the *b*-value. Therefore, we will impose a prior value taken from the region source zone and use a penalized maximum likelihood technique to fit the data (Veneziano and Van Dyck, 1985). Two options will be considered for the penalty (weight) assigned to the prior. A strong prior will produce a *b*-value very similar to the regional value with a similar degree of uncertainty. A weak prior will allow greater variation from the regional *b*-value and will result in greater uncertainty in the *b*-value for the local source. These two options are given equal weight in the hazard model. The spatial distribution of seismicity will be assumed to be uniform along the length of the local source.

Figure 5-13 shows the logic tree for the case when no source zones are defined (Plate 13) and only spatially varying seismicity rates are used to compute the hazard. The region will be assumed to have the characteristics of a single source zone and the EPRI (1994) approach will be used to assess the maximum magnitude. Seismicity parameters will be estimated using the spatially varying model developed by EPRI (1986) and a one-degree cell size.

Figure 5-14 shows the logic tree to be used for the regional seismicity zones (Plate 13) and the zones defined from the combined lineament-seismicity analysis (Plate 11). The approaches that will be used are the same as those judged appropriate for the regional source zones except that only the uniform spatial density model will be used because the zones were defined to encompass areas of uniform seismicity.

TABLE 5-1
THE SET OF DESCRIPTOR VARIABLES AND THE LEVELS OF EACH USED TO IDENTIFY LIKE DOMAINS (FROM JOHNSTON AND OTHERS, 1994)

Descriptor Variable	Levels or Values	(Base Level)*
1. Crustal Type	Rift Margin Margin* Non-extended crust	 (base)
2. Geological Age	Cenozoic Mesozoic Paleozoic Pre-Cambrian	(base)
3. Stress Regime	Compressional Extensional Unclear	 (base)
4. Angle of Stress with Structure**	Unfavorable (angle ≥25°) Favorable (angle <25°) Unknown	 (base)
Sub-Variables (Used only for Certain Crustal Types)		
5. If Crustal Type is Non-Extended Crust: Basement Age	Same Levels as above	
6. If Crustal Type is Margin, Margin*, or Rift: Cross-Structure Exists?	Yes Don't know	(base)
7. If Crustal Type is Margin, Margin*, or Rift: Is Surrounding Crust Oceanic or Continental? Side 1	Unknown Oceanic Continental	(base)
Side 2	Unknown oceanic Continental	(base)
8. If Crustal Type is Rift: Does the Domain Contain a Single or Multiple Features?	Unknown Single Multiple	(base)

* Domain that surely contains some *highly* extended region

* Default case for multivariate regression model (Table 5-2)

** This descriptor value is not included in multivariate regression model (Table 5-2)

TABLE 5-2
MULTIVARIATE REGRESSION MODEL COEFFICIENTS AND t-STATISTICS
(FROM JOHNSTON AND OTHERS, 1994)

Coefficient	Estimate	(t-statistic)
Default, b_0	6.43	
Margin, b_1	0.90	1.2
Margin*, b_2	1.20	1.5
Rift, b_3	1.07	1.4
Meso, b_4	-0.59	2.1
Paleo, b_5	-1.01	2.9
Pre-C., b_6	-0.59	1.6
Compr., b_7	0.48	2.3
Ext., b_8	0.15	0.4
Log (activity rate), b_9	0.32	1.9
Oceanic, b_{10}	-0.18	1.6
Cont., b_{11}	-1.57	2.3
$\sigma = 0.75$	$r^2 = 0.29$	

* Domain that surely contains some *highly* extended region

Regression Model:

$$M_{max} = b_o + \{b_1,b_2,b_3\}\begin{Bmatrix}x_1\\x_2\\x_3\end{Bmatrix} + \{b_4,b_5,b_6\}\begin{Bmatrix}x_4\\x_5\\x_6\end{Bmatrix} + \{b_7,b_8\}\begin{Bmatrix}x_7\\x_8\end{Bmatrix} + b_9x_9 + \{b_{10},b_{11}\}\begin{Bmatrix}x_{10}\\x_{11}\end{Bmatrix}$$

in which: $\{x_1,x_2,x_3\}$ are the zero-one indicator variables of crustal type: Margin, Margin*, and Rift; $\{x_4,x_5,x_6\}$ relate to age Mesozoic, Paleozoic, and Pre-Cambrian (Cenozoic is the default case); $\{x_7,x_8\}$ indicate compressional or extensional stress regime (unknown being the default case); x_9 is log activity rate (mean annual rate of events equal to or greater than magnitude 5 per year per 10^6 km^2; and $\{x_{10},x_{11}\}$ relate to oceanic or continental surrounding crust "on side 1" (assumed zero unless the crustal type is one of the extended types).

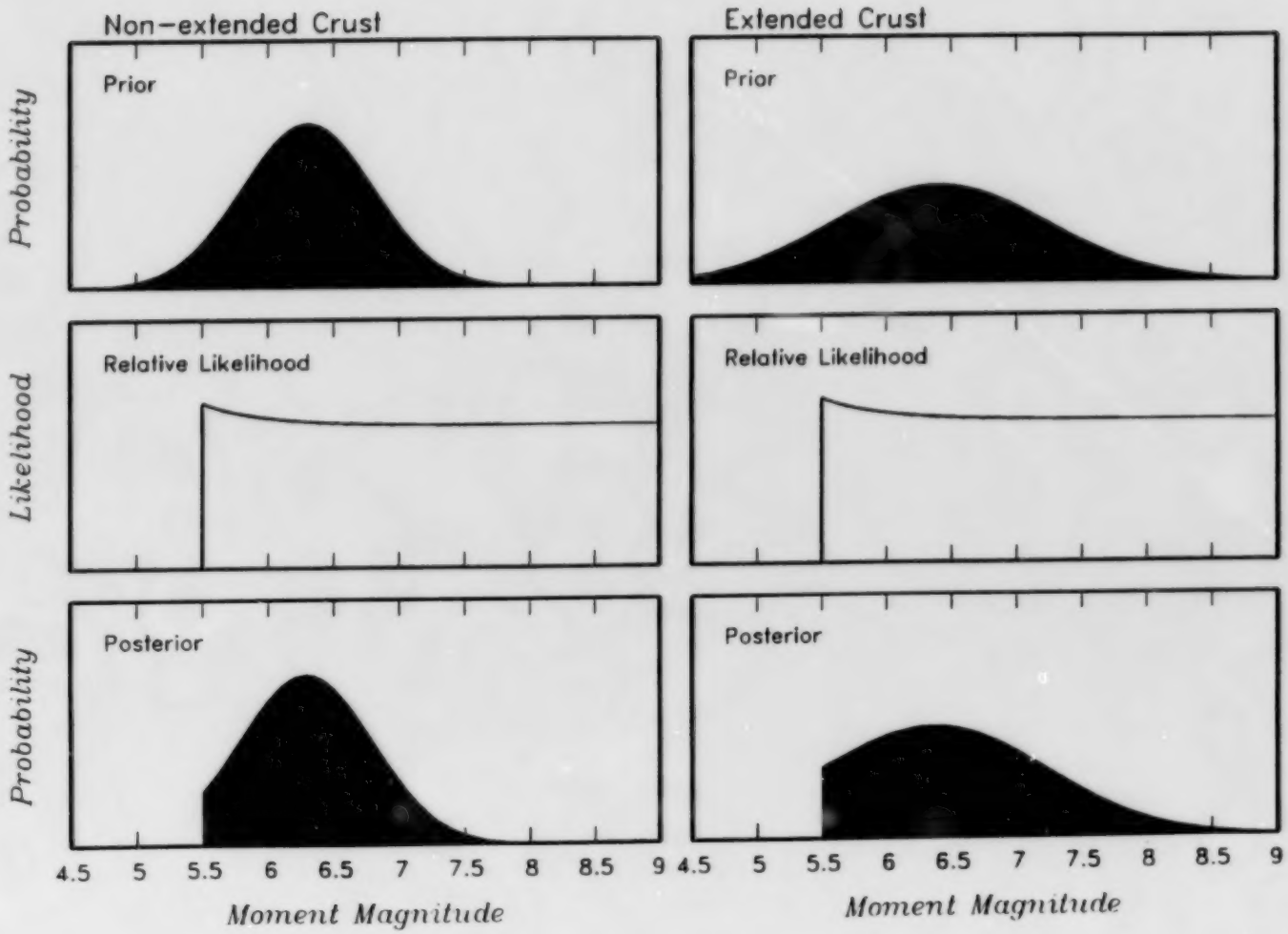


Figure 5-1 Example prior, sample likelihood, and posterior distributions for zones with one observed event of magnitude 5.5 (from Johnston and others, 1994)

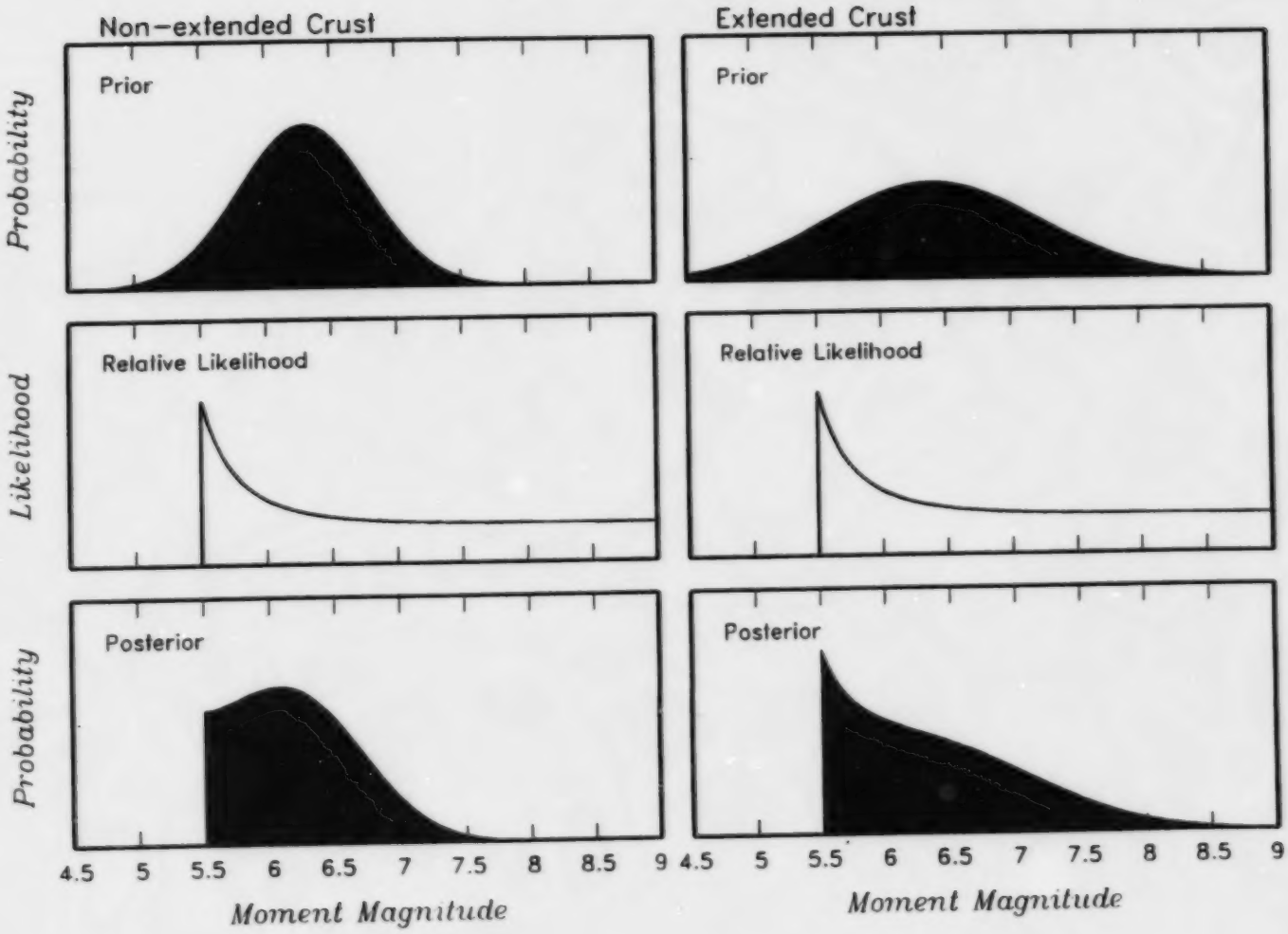


Figure 5-2 Example prior, sample likelihood, and posterior distributions for zones with 10 observed events >4.5 and largest of magnitude 5.5 (from Johnston and others, 1994)

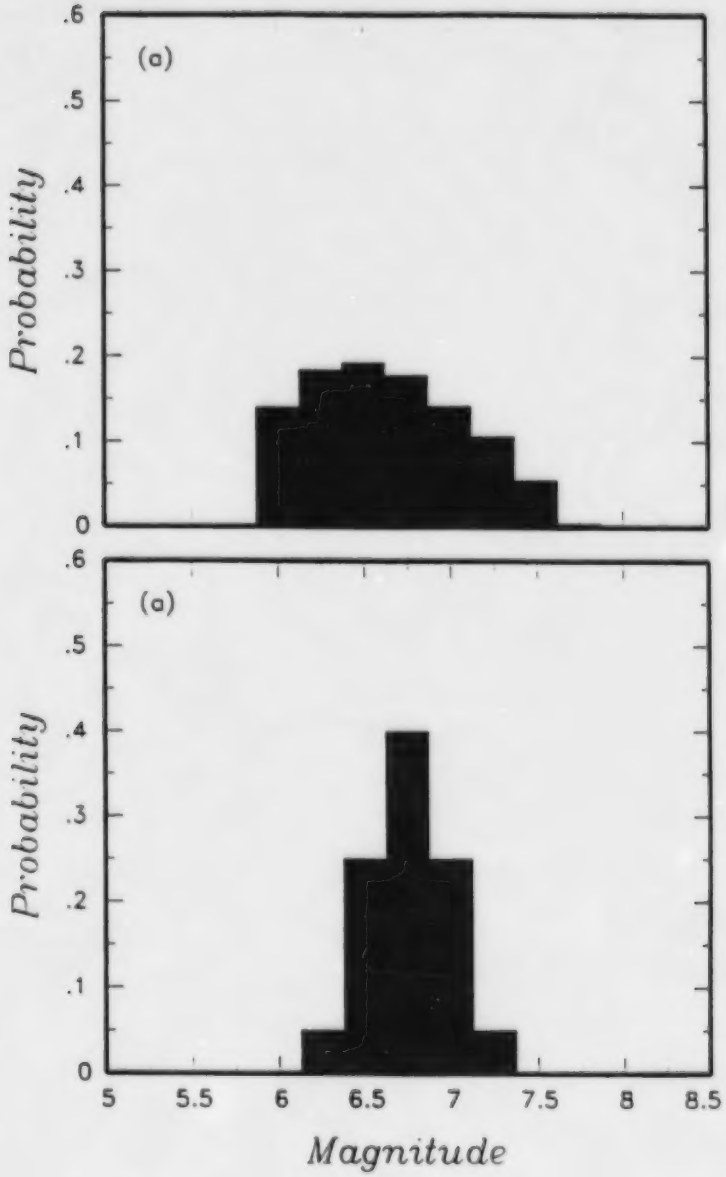


Figure 5-3 Example maximum magnitude distributions (mb) for (a) a hypothetical seismic source zone and (b) for a hypothetical local source considered to be a fault.

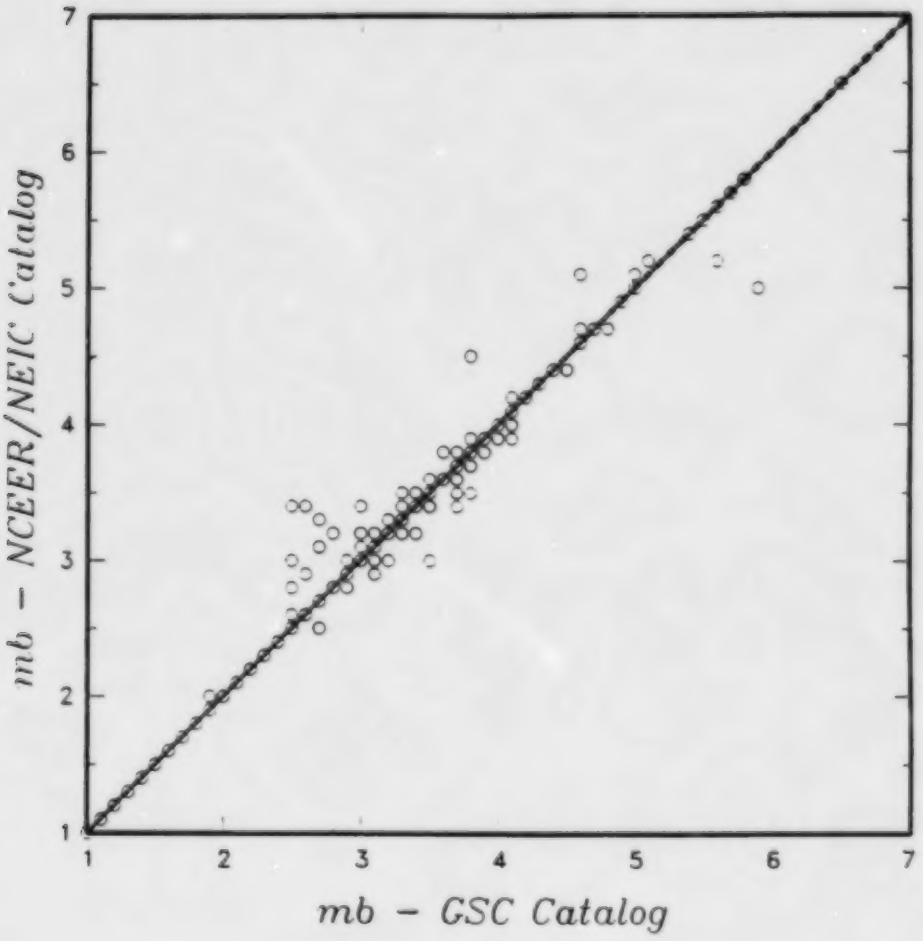


Figure 5-4 Comparison of magnitude estimates in the GSC and NCEER/NEIC catalogs. Solid line shows best fit linear regression.

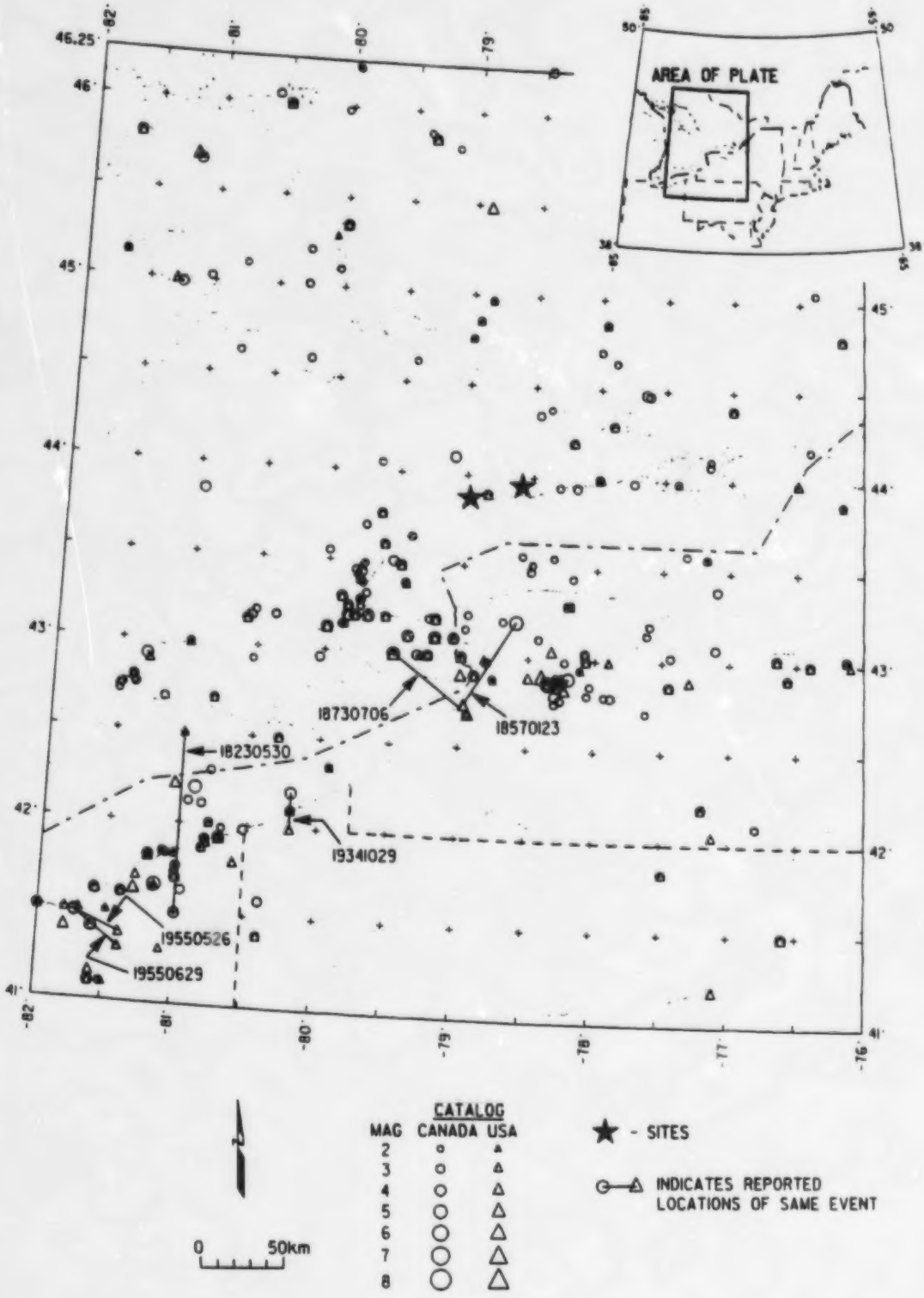


Figure 5-5 Local seismicity (from Canadian and United States catalogs)

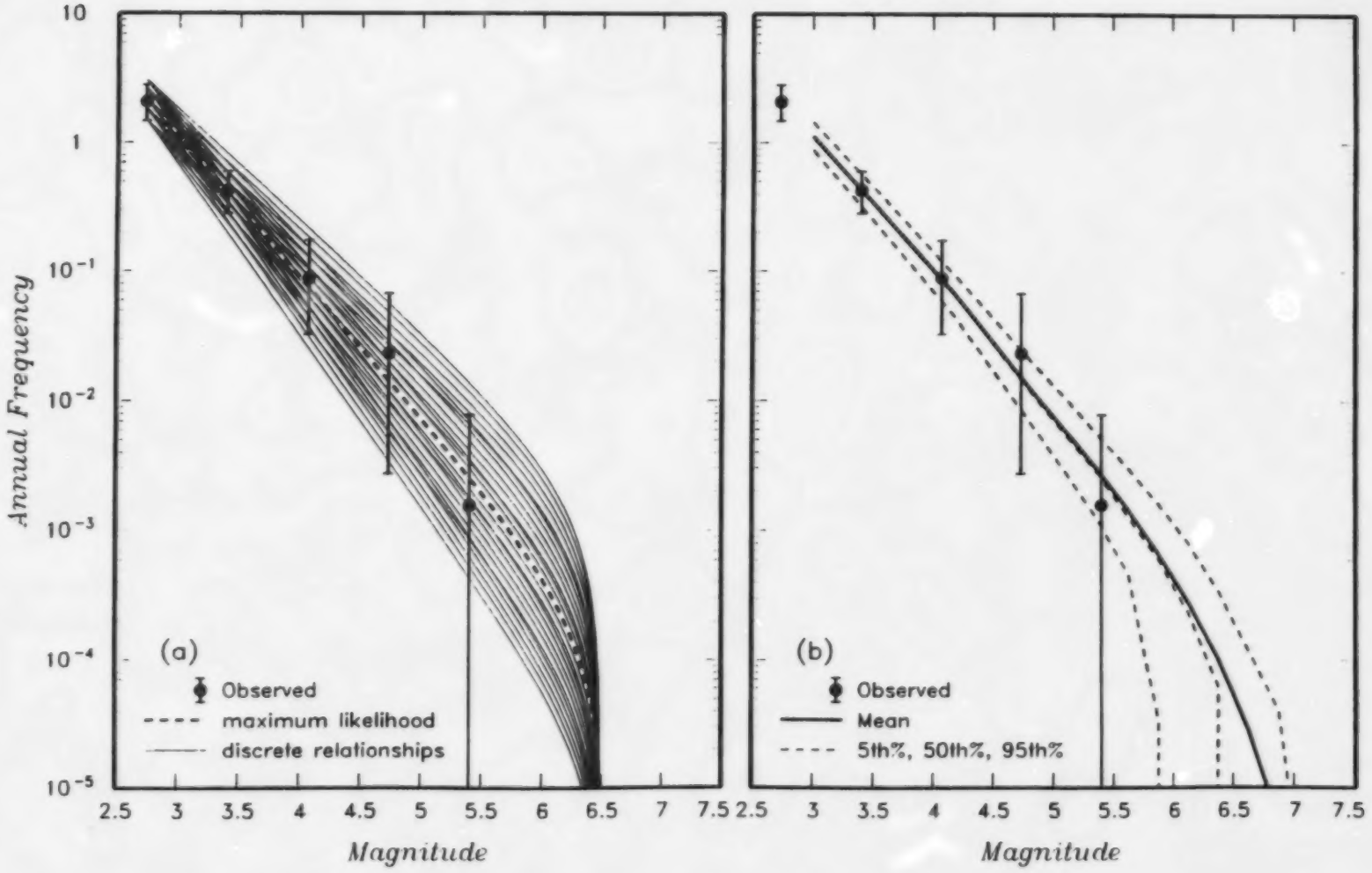


Figure 5–6 Example of uncertainty assessment for source zone recurrence parameters (a) discrete recurrence relationships fitted to observed data for Mmax of 6.5. (b) Resulting uncertainty in recurrence rates including uncertainty in Mmax.

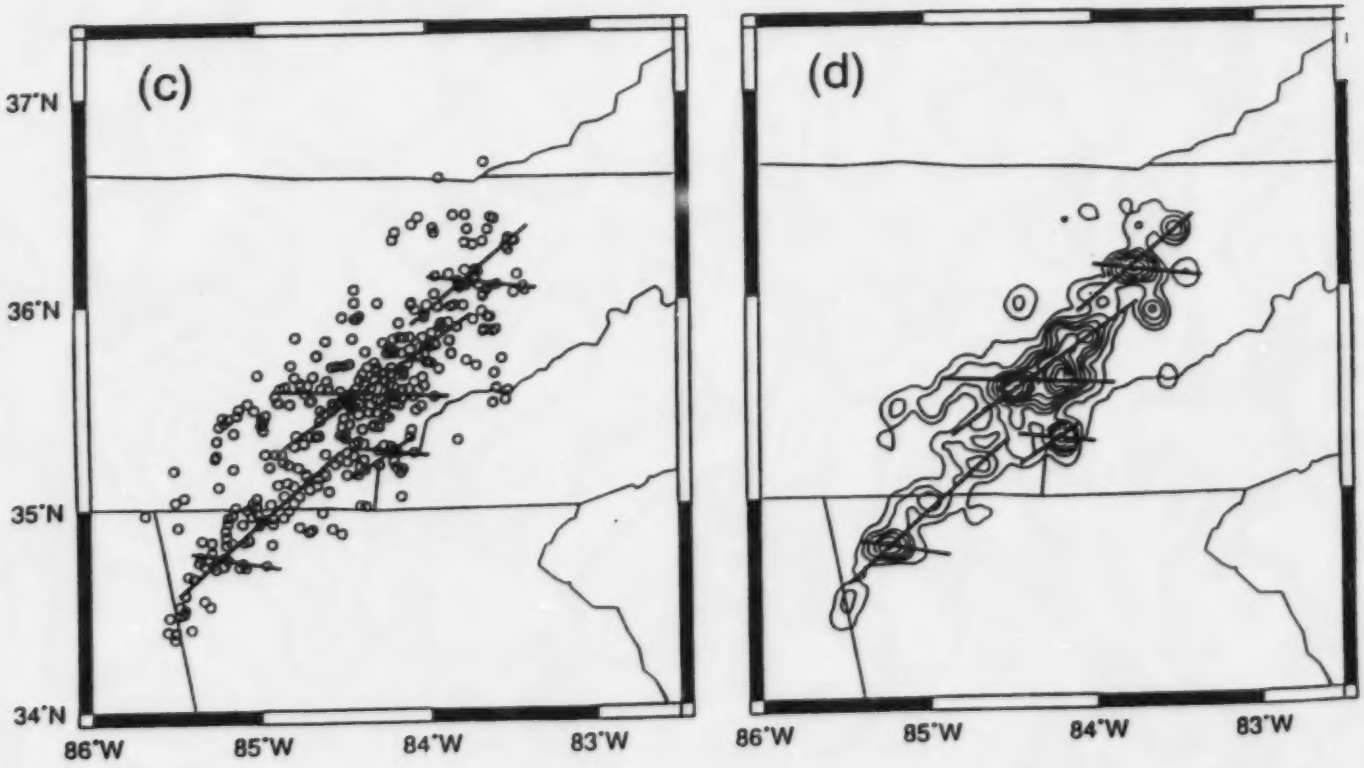


Figure 5-7 Example of spatial density estimate for seismicity rate obtained by kernel density estimation. (c) Spatial distribution of epicenters. (d) resulting spatial density function. (from Chapman and others, 1996 in press.)

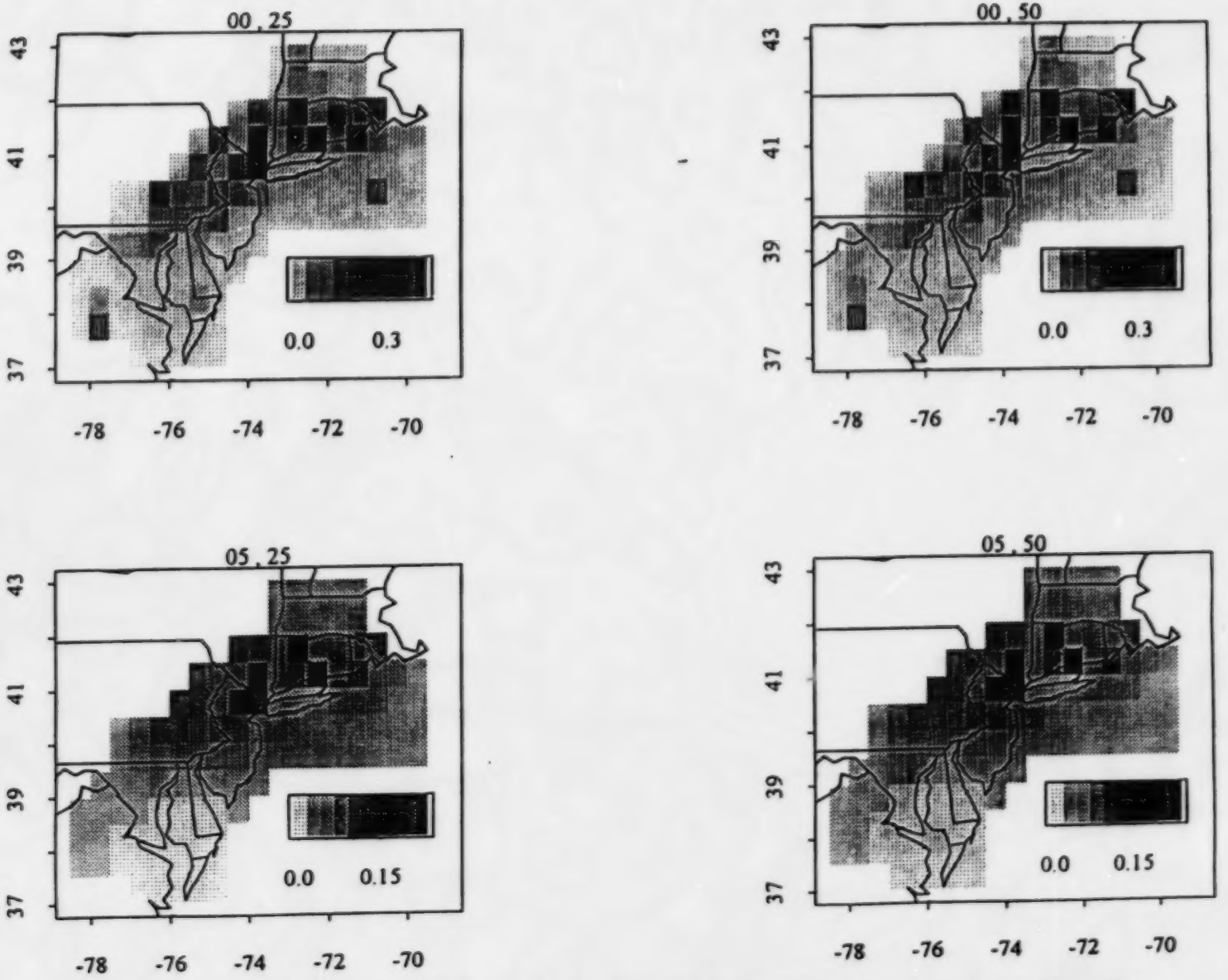


Figure 5-8 Spatial distribution of seismicity rate for an eastern coastal margin regional source zone. The numbers above each plot indicate the penalties placed on the rate and *b*-value parameters (from Geomatrix, 1995)

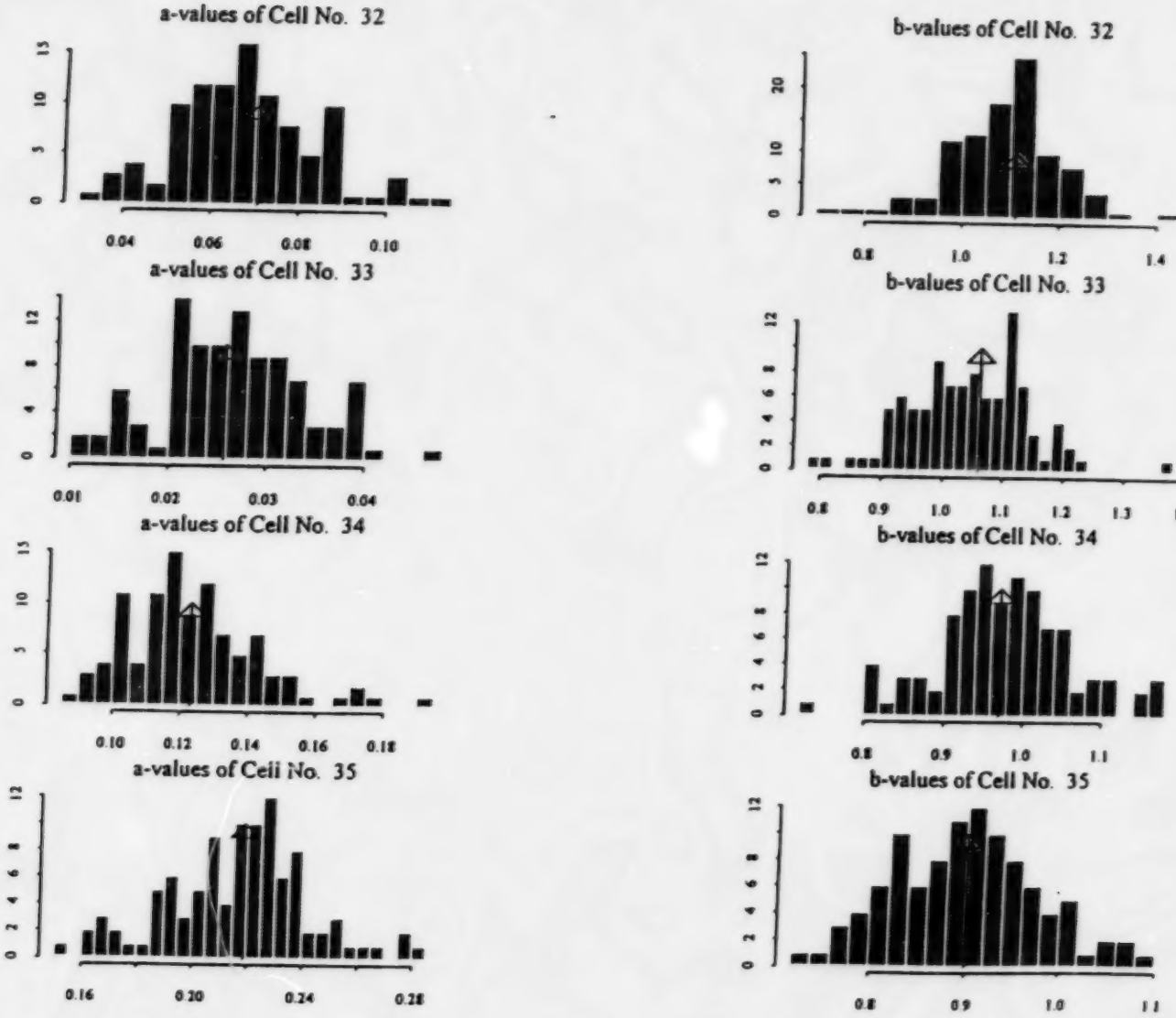


Figure 5-9 Histograms showing the distribution of 100 simulations of rate and *b*-value for four subzones of the regional source zone shown on Figure 5-8. Arrows indicate the maximum likelihood solution for the subzone (from Geomatrix, 1995)

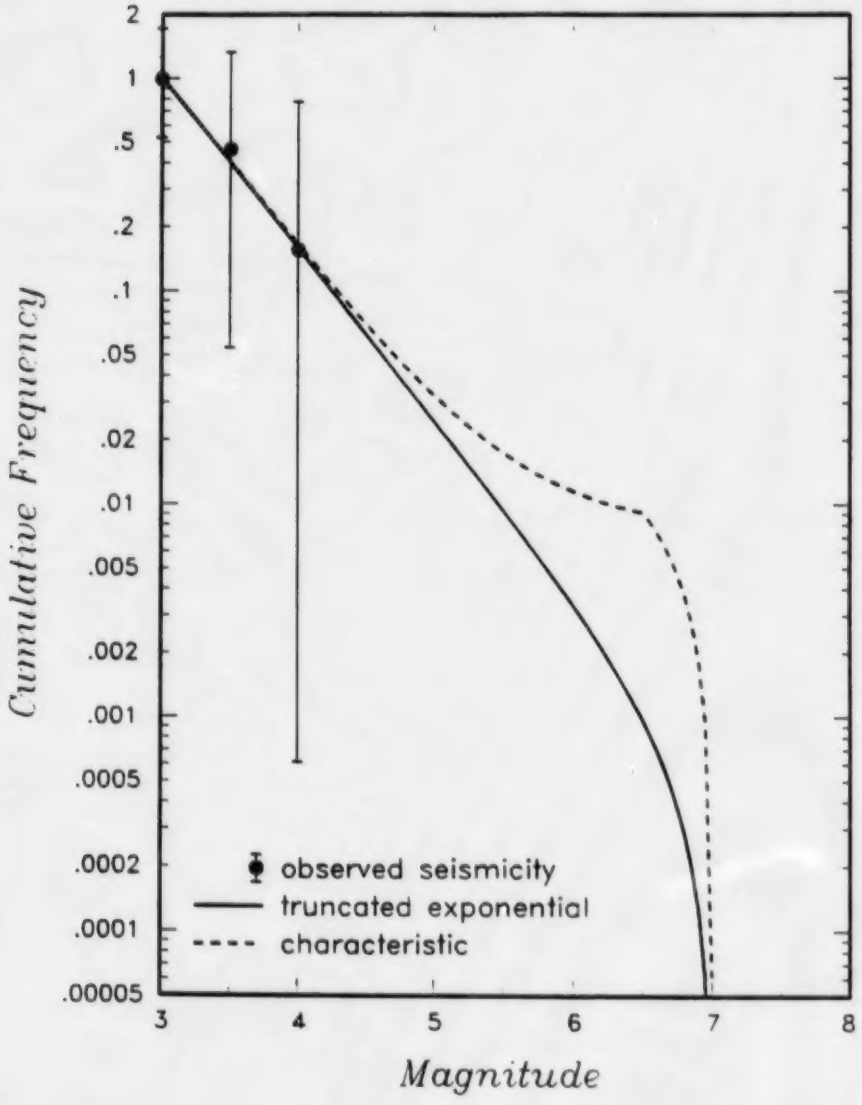


Figure 5-10 Magnitude distribution models fit to observed seismicity.

Source Type	*Source Model*	*Maximum Magnitude Approach*	*Maximum Magnitudes*	*Seismicity Rate*	*b-value*	*Magnitude Distribution*	*Spatial Distribution*

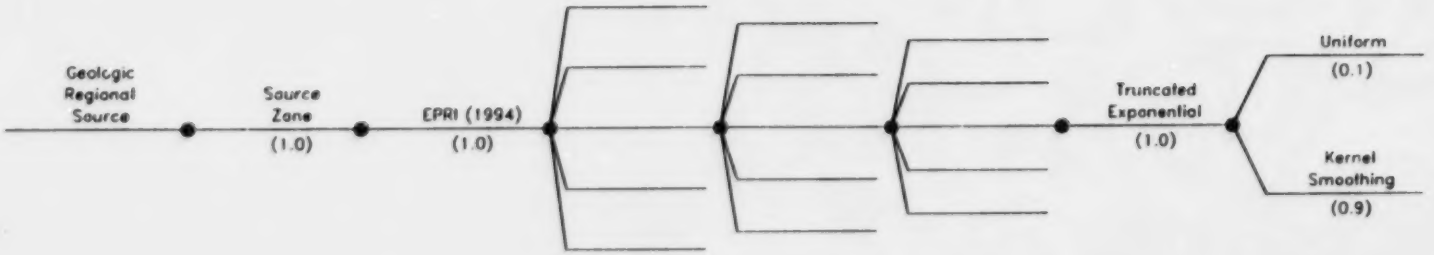


Figure 5-11 Seismicity parameter logic tree for geologically defined regional source zones (Plates 1 through 10)

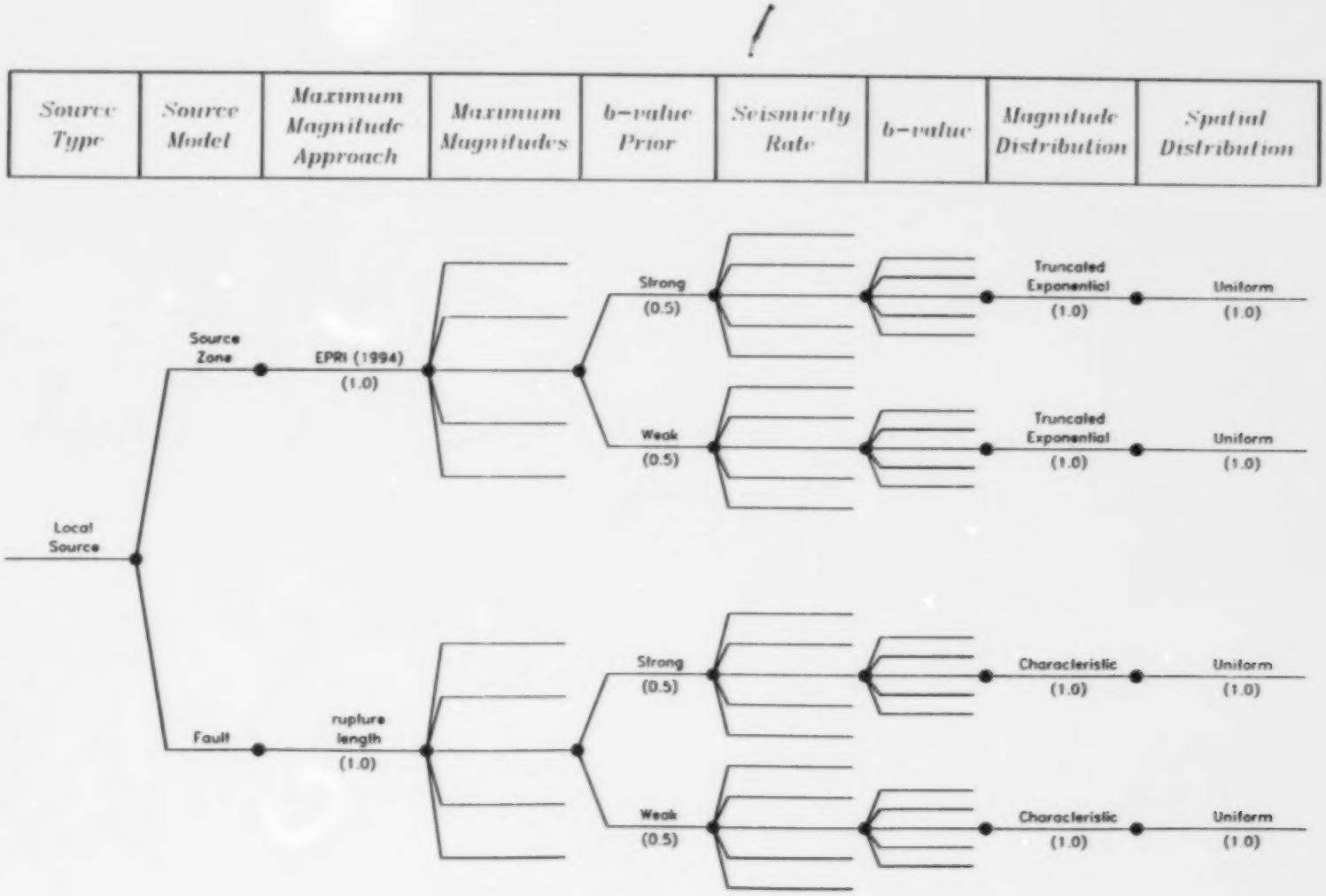


Figure 5-12 Seismicity parameter logic tree for local sources (Plates 14 and 15)

Source Type	*Source Model*	*Maximum Magnitude Approach*	*Maximum Magnitudes*	*Joint Seismicity Rate and b-value*	*Magnitude Distribution*	*Spatial Distribution*

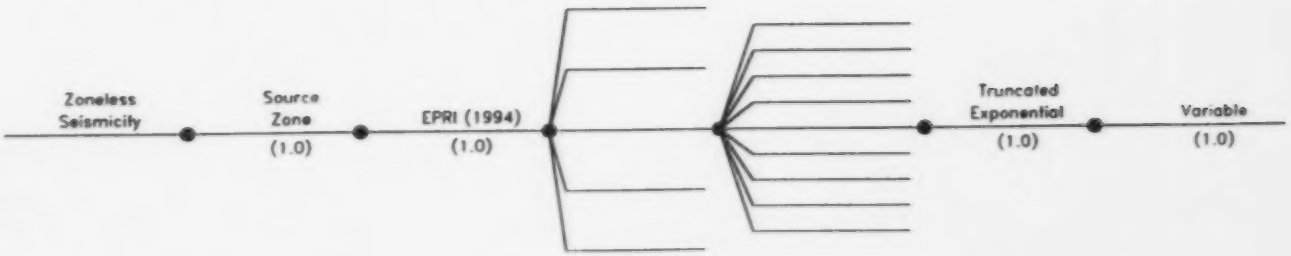


Figure 5-13 Seismicity parameter logic tree for zoneless seismicity model (Plate 12)

Source Type	*Source Model*	*Maximum Magnitude Approach*	*Maximum Magnitudes*	*Seismicity Rate*	*b-value*	*Magnitude Distribution*	*Spatial Distribution*

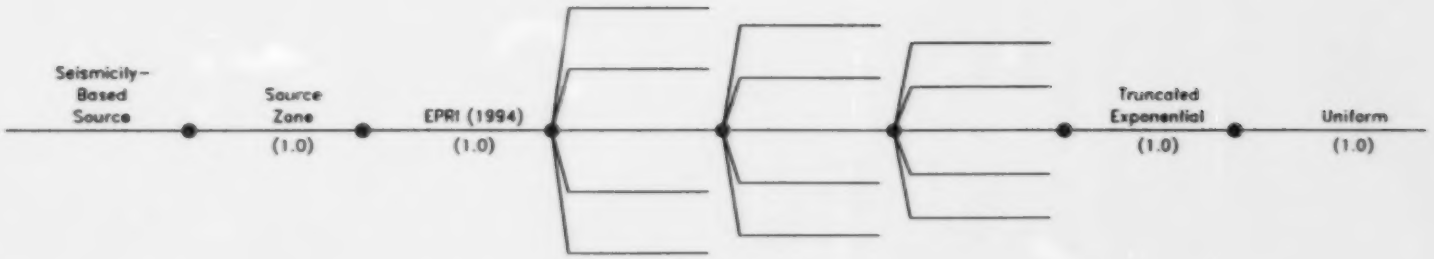


Figure 5-14 Seismicity parameter logic tree for seismicity-based source zones (Plates 11 and 13)

6.0 GROUND MOTION ATTENUATION MODELS

6.1 SUMMARY OF RECENT WORK

In the past decade there has been substantial research effort focused on developing strong ground motion models for the stable continental regions of eastern North America (ENA). Because of the very limited strong motion data base, these models have been developed from analysis of seismographic network data (e.g. Atkinson and Mereu, 1992) and the use of numerical ground motion models (e.g. Atkinson, 1984; Burger and others, 1987; Boore and Atkinson, 1987; Toro and McGuire, 1987; Somerville and others, 1990; Atkinson and Boore, 1990, 1995; Boore and Joyner, 1991; EPRI, 1993, Toro and others, 1997). These studies are summarized in Atkinson and Boore (1997).

Examination of the most recent models, those of EPRI (1993, see also Toro and others, 1997) and Atkinson and Boore (1995), indicates two points where significant differences occur. First, the two models use different representations of the earthquake source spectrum. Atkinson and Boore (1995) used the empirical source model of Atkinson (1993) while EPRI (1993) used the more common Brune (1970, 1991) model. These two alternative source models have a significant impact on the level and relative frequency content of ground motions in the frequency range of importance to nuclear power plants. Figure 6-1 compares the assessments for peak ground acceleration (PGA) and 5-percent damped spectral acceleration (SA) at frequencies of 10 Hz and 1 Hz developed by EPRI (1993) and Atkinson and Boore (1995). The EPRI (1993) model uses surface distance to the rupture (asperity) as the distance measure while Atkinson and Boore (1995) used distance to the asperity. The results for Atkinson and Boore (1995) shown on Figure 6-1 were adjusted to surface distance by assuming an average depth equal to the depth parameter (C7) of the EPRI (1993) attenuation relationships (9.3, 8.5, and 6.8 km for PGA, 10 Hz SA, and 1 Hz SA, respectively). The two models predict similar estimates of high frequency motion. However, the earthquake source model used by Atkinson and Boore (1995) predicts much lower 1Hz motions than the EPRI (1993) model. Recently, Haddon (1996) proposed a third earthquake source model based on analyses of recordings from the Saguenay and Mont-Laurier earthquakes. Haddon's model suggest higher high-frequency motions than those obtained using the EPRI (1993) and

Atkinson and Boore (1995) models and comparable long period motions to those obtained from the EPRI (1993) model.

The second difference between the two attenuation models is that EPRI (1993) developed a detailed assessment of the epistemic uncertainty in the median ground motion attenuation relationship based on uncertainties in the basic modeling parameters. This estimate of epistemic uncertainty was intended to be used to construct multiple attenuation relationships for use in PSHA's. However, the modeling of the epistemic uncertainty did not include consideration of alternative earthquake source spectral shapes. The Atkinson and Boore (1995) model does provide the effect of an alternative source spectra shape but they do not provide a detailed estimate of the epistemic uncertainty in their attenuation model, though an approximate estimate is given in Atkinson and Boore (1997).

Recently, SSHAC(1996) conducted a demonstration expert elicitation of appropriate ground motion models for ENA. Important issues addressed in this effort were the shape of earthquake source spectra, scaling to large magnitudes, and the qualification of variability or randomness of ground motions and the uncertainty in attenuation modeling in ENA. Although the SSHAC (1996) effort was intended to provide a demonstration of the elicitation process rather than a definitive assessment, nevertheless, the study provides valuable guidance on the current state of knowledge for characterizing ENA strong ground motion for seismic hazard studies. The SSHAC (1996) study presented an assessment of ground motions from ENA earthquakes that addresses the median levels and the epistemic uncertainty in the median levels. These assessments were made by a panel of ground motion experts that were very familiar with the results of the EPRI (1993) study and the model proposed by Atkinson and Boore (1995). The SSHAC (1996) results are in the form of assessments of ground motions for a limited number of discrete magnitude-distance pairs and are shown on Figure 6-1. The published SSHAC results were adjusted to surface distance following the approach outlined above for Atkinson and Boore (1995). The median estimates lie between those obtained from EPRI (1993) and the Atkinson and Boore (1995) simulation results. [Note the curves labeled Atkinson and Boore (1995) on Figure 6-1 are their published parametric fit to the simulation results.]

The comparison of the EPRI (1993), SSHAC (1996), andAtkinson and Boore (1997) estimates of epistemic uncertainty are as follows:

			Epistemic Standard Error		
Frequency	m_{blg}	Distance	SSHAC (1996)	EPRI (1993)	Atkinson and Boore (1997)
1 Hz	5.5	20	0.48	0.30	≤ 0.7
		70	0.46	0.30	≤ 0.7
		200	0.37	0.30	≤ 0.7
	7.0	20	0.66	0.54	≤ 0.7
		70	0.71	0.54	≤ 0.7
		200	0.73	0.54	≤ 0.7
10 Hz	5.5	20	0.32	0.29	≤ 0.7
		70	0.32	0.29	≤ 0.7
	7.0	20	0.52	0.45	≤ 0.7
		70	0.53	0.45	≤ 0.7
PGA	5.5	70	0.41	0.29	≤ 0.7
	7.0	70	0.51	0.45	≤ 0.7

The EPRI (1993) estimates of the epistemic uncertainty in the median attenuation relationships are lower that the SSHAC (1996) estimates. This is likely due to consideration of only one earthquake source model in the development of the EPRI (1993) relationships. The Atkinson and Boore (1997) rough estimate is larger than both the EPRI (1993) and SSHAC (1996) values.

The comparison of the EPRI (1993), Atkinson and Boore (1995), and SSHAC (1996) estimates of aleatory uncertainty (randomness) are as follows:

			Aleatory Standard Error		
Frequency	m_{blg}	Distance	SSHAC (1996)	EPRI (1993)	Atkinson and Boore (1995)
1 Hz	5.5	20	0.80	0.91	0.99
		70	0.80	0.91	0.99
		200	0.80	0.91	0.99
	7.0	20	0.78	0.80	0.99
		70	0.78	0.80	0.99
		200	0.79	0.80	0.99
10 Hz	5.5	20	0.73	0.62	0.67
		70	0.75	0.62	0.67
	7.0	20	0.70	0.51	0.67
		70	0.72	0.51	0.67
PGA	5.5	70	0.75	0.62	0.67
	7.0	70	0.70	0.52	0.67

The EPRI (1993) and SSHAC (1996) estimates of aleatory uncertainty are comparable at low frequencies and the SSHAC (1996) estimates are significantly higher at high frequencies. However, SSHAC (1996) points out that their estimates for high frequency aleatory uncertainty may be too high. The Atkinson and Boore (1995) estimates are similar to the EPRI (1993) for m_{bLg} 5.5 and are larger for m_{bLg} 7.

6.2 Ground Motion Modeling Approach

The focus of this study is quantifying the important issues in seismic source characterization, including quantifying the uncertainty in the seismic hazard due to the scientific uncertainty in identifying and characterizing seismic sources. Thus, it is important to properly characterize the uncertainty in ground motion modeling so that the relative significance of all the uncertainties in the seismic hazard can be identified.

Two approaches can be used to address the epistemic uncertainty in ground motion modeling. The most common approach is the use of multiple attenuation relationships selected as representative of the current state of scientific knowledge. An alternative would be to use the results of the SSHAC (1996) study to describe the expected ground motion levels for the specific magnitudes and distances that they considered. Interpolation to other magnitudes and distances can be achieved by fitting an attenuation model to the specific data presented. The first approach provides a clear identification of the results with a specific attenuation relationship while the second approach may give a fuller expression of the epistemic uncertainty that may not be captured in a specific attenuation model.

The introduction of a third source spectra model by Haddon (1996) that was not explicitly considered in SSHAC (1996) suggests that the approach for this study is to use multiple attenuation models. We intend to use the median attenuation relationships developed by EPRI (1993) and the median simulation results developed by Atkinson and Boore (1995). In order to develop attenuation relationships for the Atkinson and Boore (1995) model that are at the same spectral frequencies as those used by EPRI (1993) the simulations will be repeated and a functional form that more closely follows the simulation model will be fit to the results. A

third ground motion attenuation model will be constructed using the source spectra model developed by Haddon (1996) in the simulation algorithm used by Atkinson and Boore (1995).

The EPRI (1993) model provides an assessment of the epistemic uncertainty resulting from uncertainties in all the model parameters except the shape of the source spectra. We intend to use the EPRI (1993) estimate of epistemic uncertainty to characterize the epistemic uncertainty in each of the three attenuation models. The approach outlined in EPRI (1993) is to replace normal distribution modeling the epistemic uncertainty in the log of peak motion by a discrete set of attenuation relationships. EPRI (1993) used a four point distribution that matches the first six moments of a standard normal distribution. In this study we propose to use the three point distribution defined above (Keefer and Bodily, 1983) because we are also using three attenuation relationships to represent uncertainty in the shape of the source spectrum, resulting in nine total relationships. Figure 6-2 shows the logic tree for ground motion model uncertainty with the specific values of standard normal deviation and their associated weights. We give the highest weight (0.6) to the EPRI (1993) model because it was developed by a group of experts and uses a more widely accepted source model. The model of Atkinson and Boore (1995) is given a weight of 0.3 reflecting the fact that it uses a less widely accepted source model. The attenuation model based on the Haddon (1996) source spectra is given the lowest weight (0.1), reflecting the limited data on which it is based. As indicated on Figure 6-2, and Figure 1-2, the seismic source characterization assessments (presented by the logic trees in Section 4 and 5) follow the attenuation uncertainty characterization.

The EPRI (1993), SSHAC (1996), and Atkinson and Boore (1995) estimates of aleatory uncertainty are comparable at low frequencies and the SSHAC (1996) estimates are significantly higher at high frequencies. As indicated above, the SSHAC (1996) estimates for aleatory uncertainty may be too high. Therefore, we intend to use the EPRI (1993) model for aleatory uncertainty for each of the nine attenuation models defined on Figure 6-2.

The selected attenuation models are all defined for hard rock site conditions. Because the interest of the study is in the analysis of the uncertainty in the hazard rather than the absolute hazard levels, it is not necessary to correct the results for local site conditions.

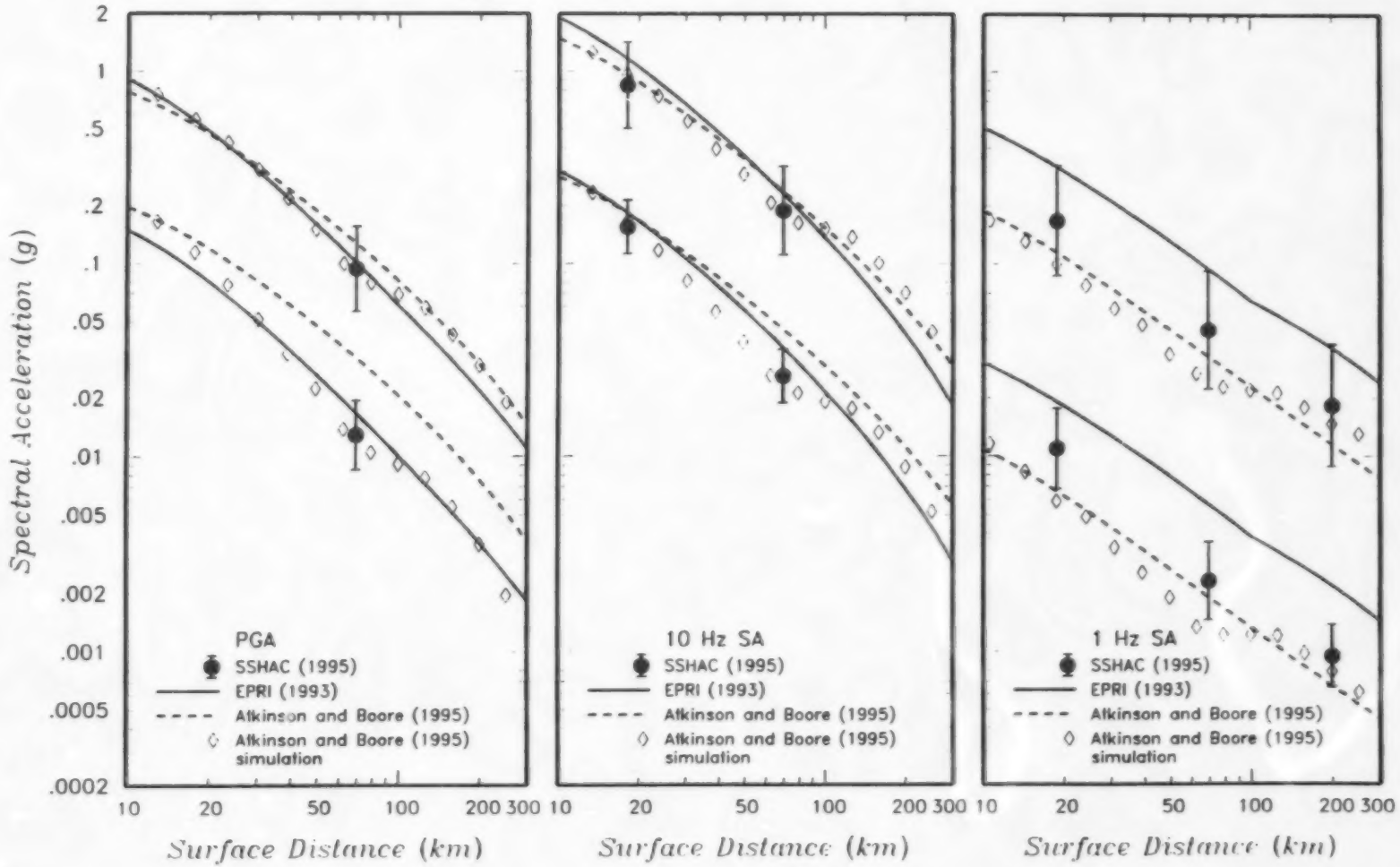


Figure 6–1 Comparison of median ground motions from SSHAC (1995), EPRI (1993), and Atkinson and Boore (1995). Vertical bars denote episermic uncertainty in SSHAC (1995)

Attenuation Model	*Normalized Attenuation Epistemic Standard Error*	*Seismic Source Characterization Assessments*

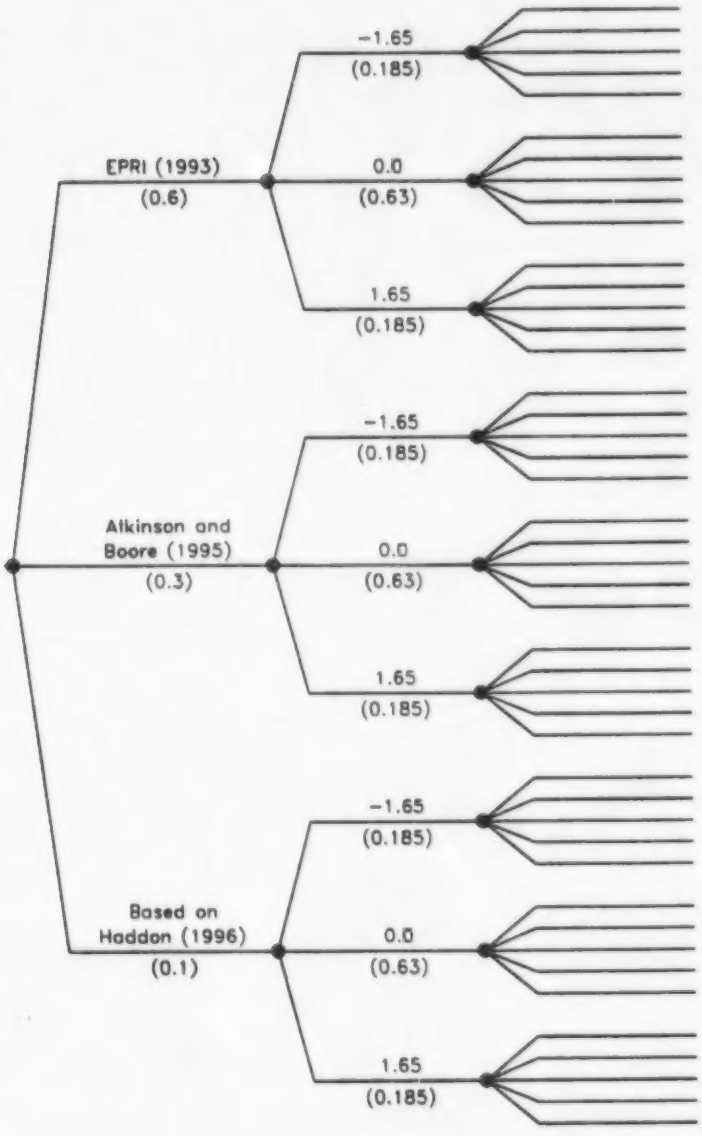


Figure 6–2 Logic tree for ground motion assessment.

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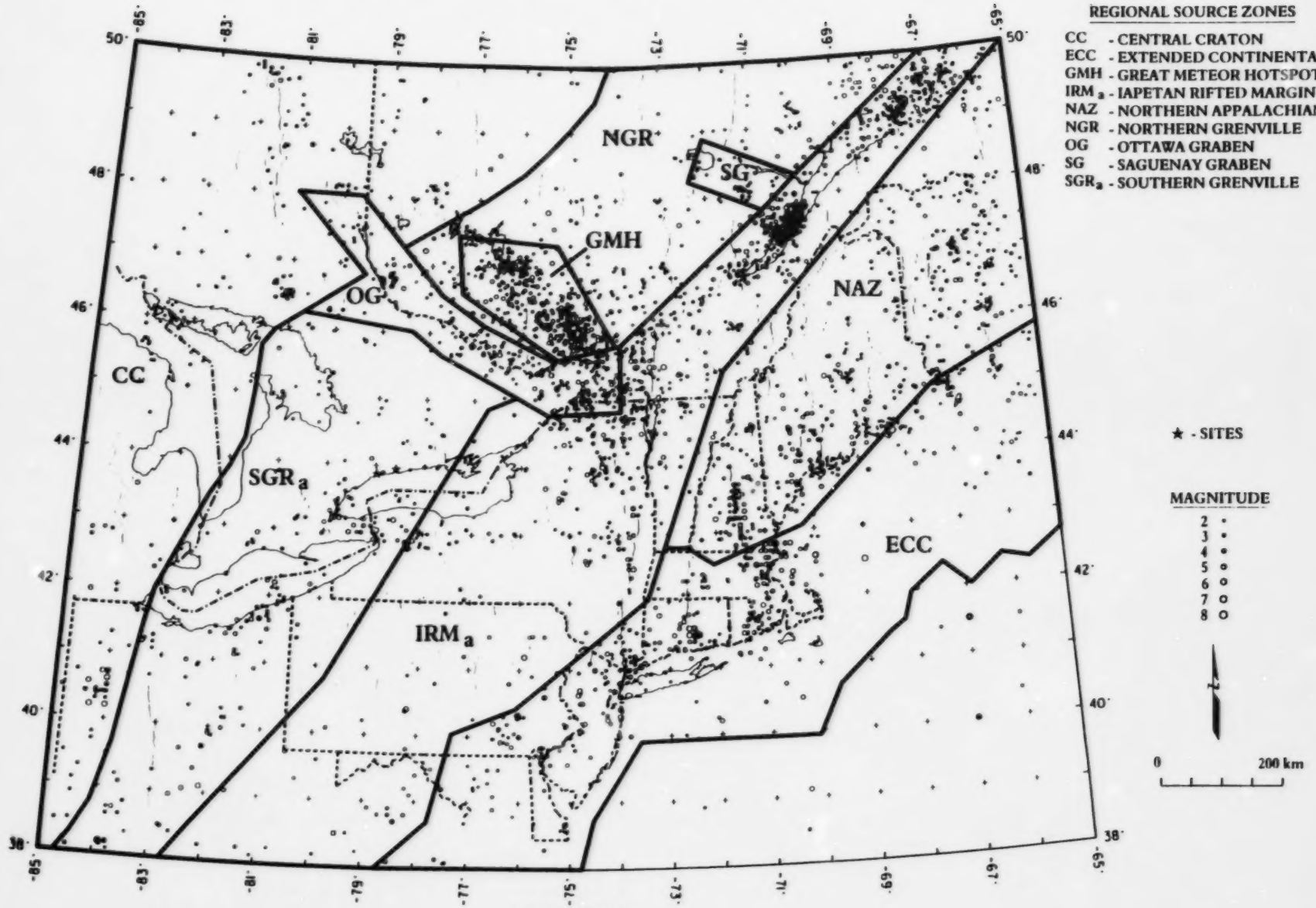

PLATE 1A

REGIONAL SOURCE ZONES FOR WHEELER IRM/SGR BOUNDARY

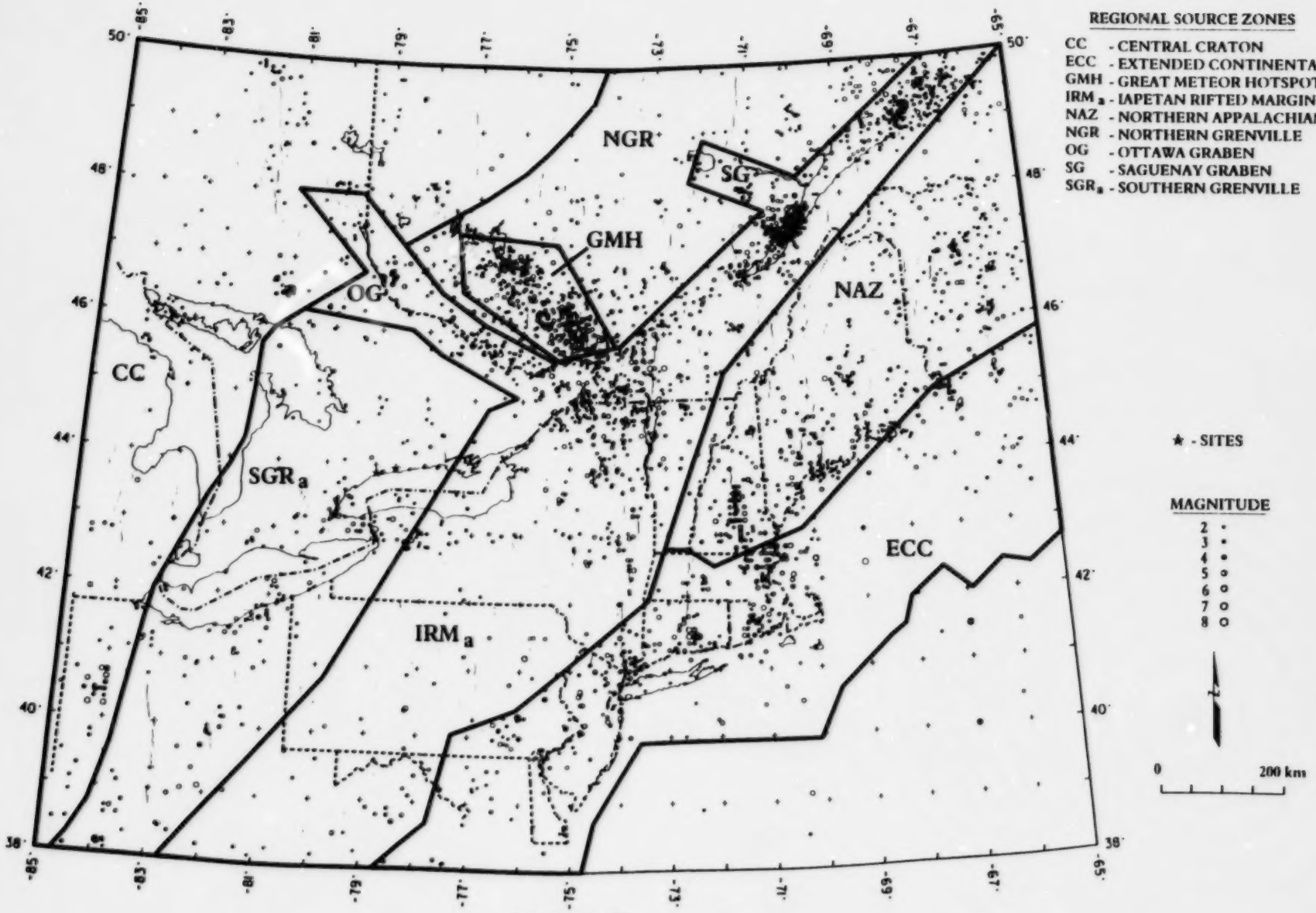

PLATE 1B

REGIONAL SOURCE ZONES FOR WHEELER IRM/SGR BOUNDARY

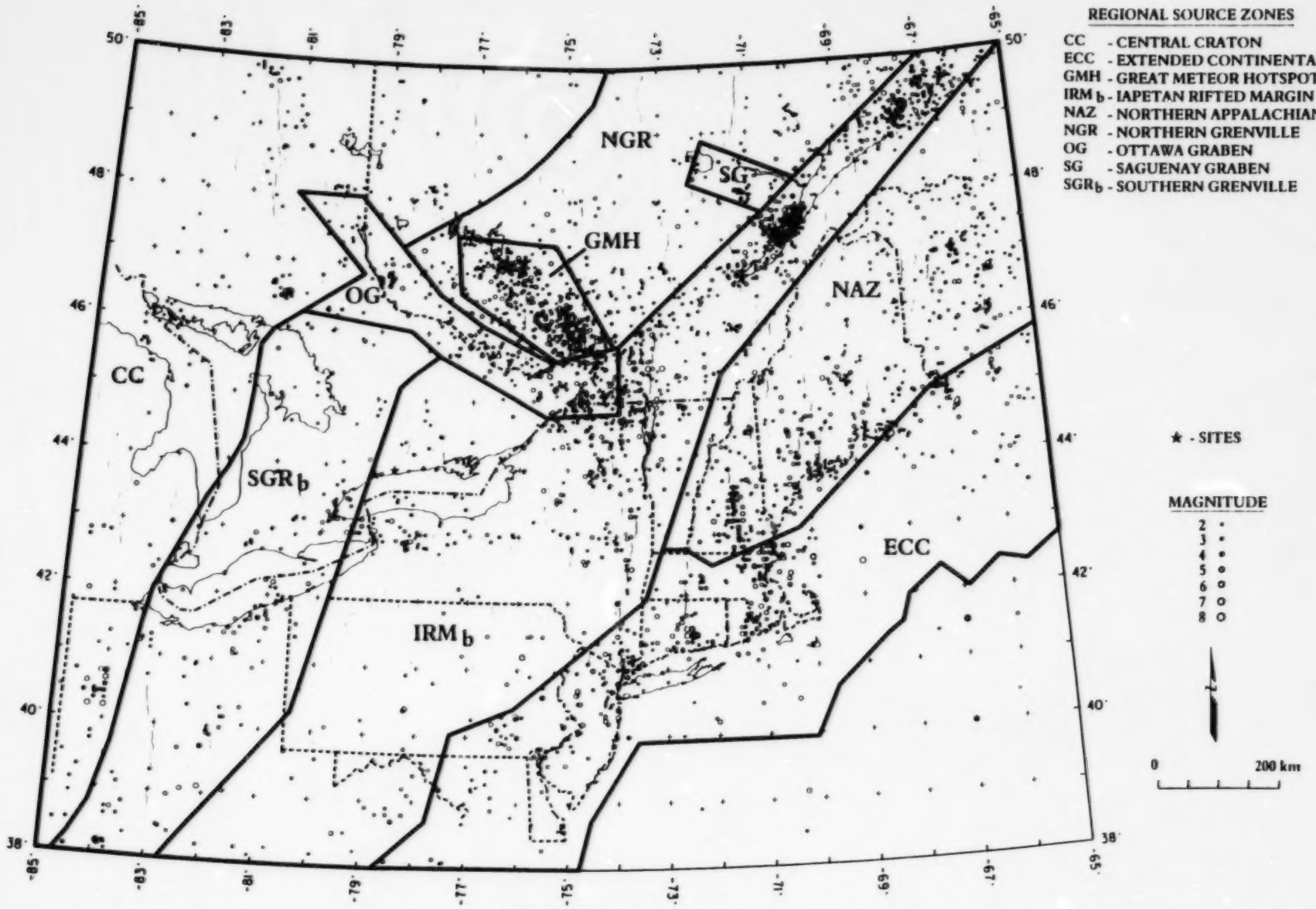

PLATE 2A

REGIONAL SOURCE ZONES FOR CMBBZ IRM/SGR BOUNDARY

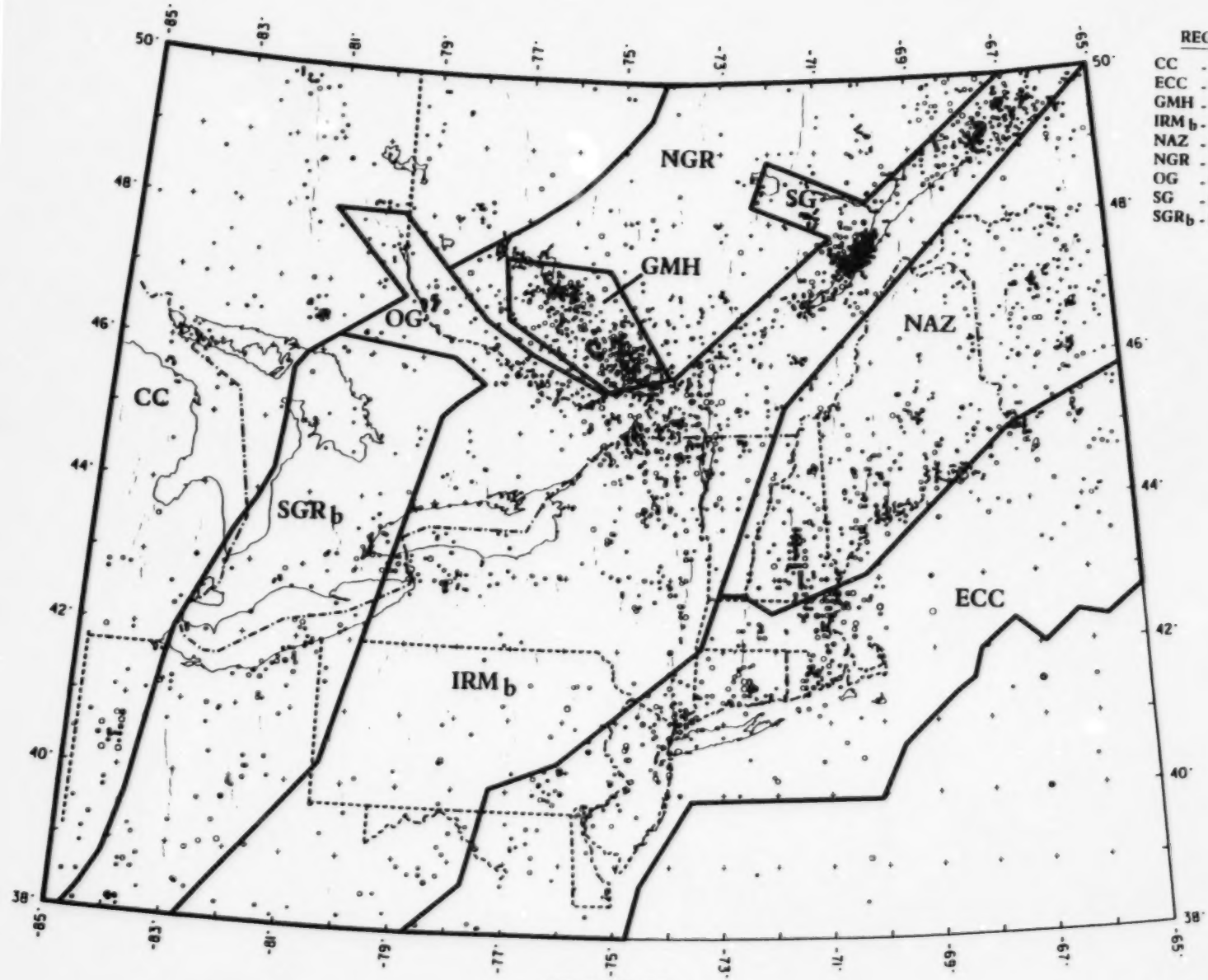

PLATE 2B

REGIONAL SOURCE ZONES FOR CMBBZ IRM/SGR BOUNDARY

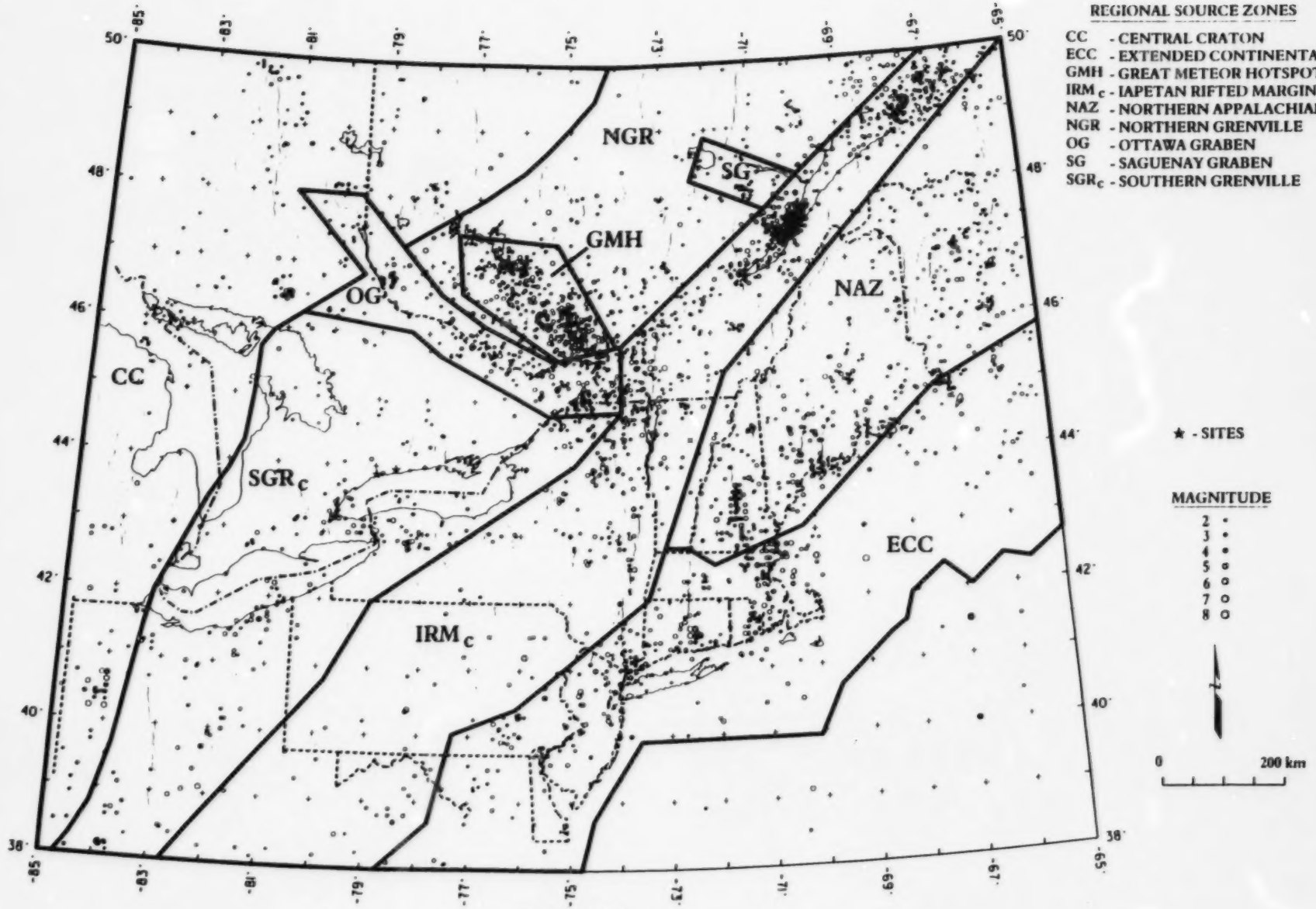

PLATE 3A

REGIONAL SOURCE ZONES FOR WESTERN ADIRONDACK UPLIFT IRM/SGR BOUNDARY

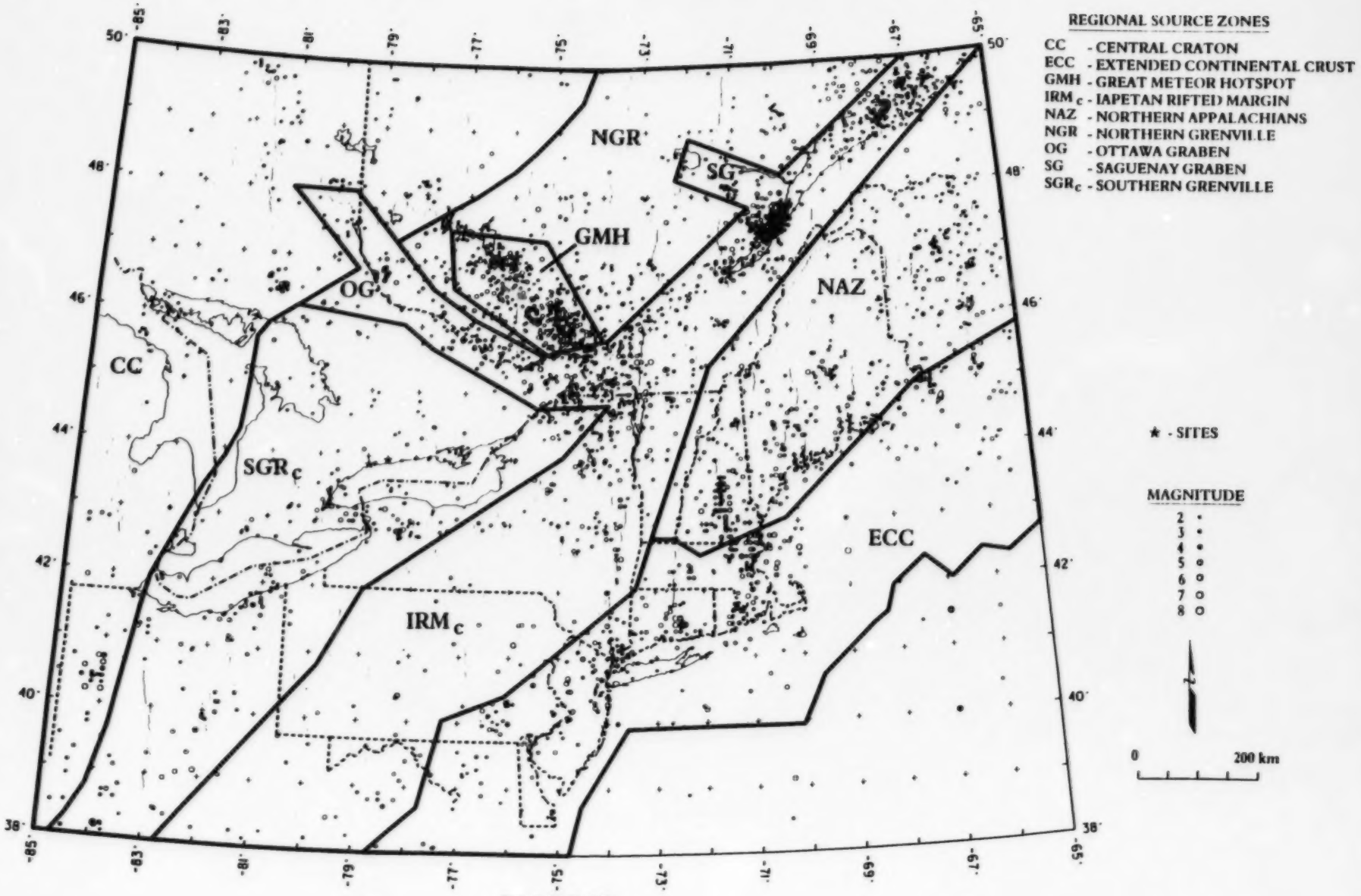

PLATE 3B

REGIONAL SOURCE ZONES FOR WESTERN ADIRONDACK UPLIFT IRM/SGR BOUNDARY

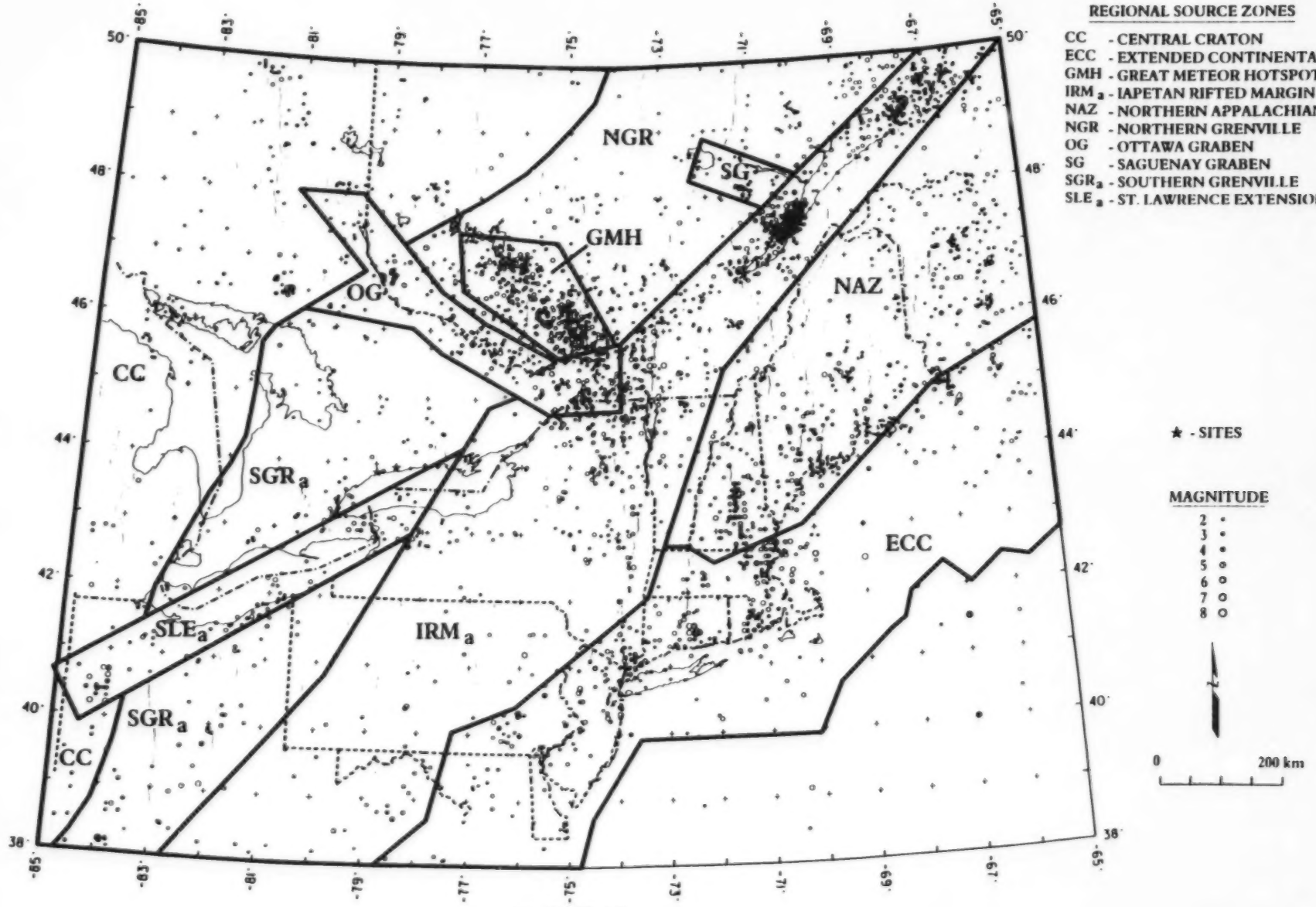

PLATE 4A

REGIONAL SOURCE ZONES FOR WHEELER IRM/SGR BOUNDARY WITH NARROW ST. LAWRENCE EXTENSION

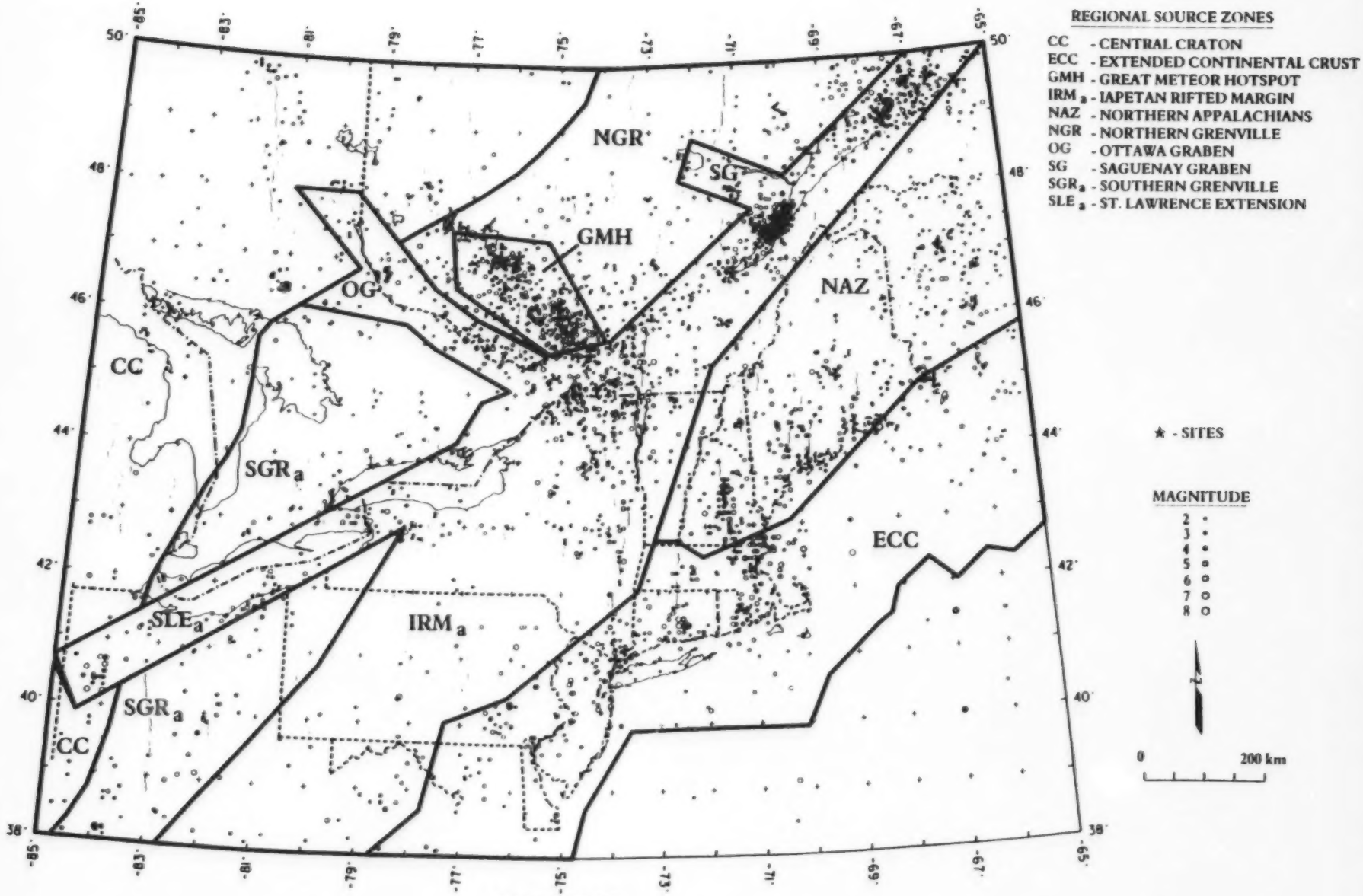

PLATE 4B

REGIONAL SOURCE ZONES FOR WHEELER IRM/SGR BUONDARY WITH NARROW ST. LAWRENCE EXTENSION

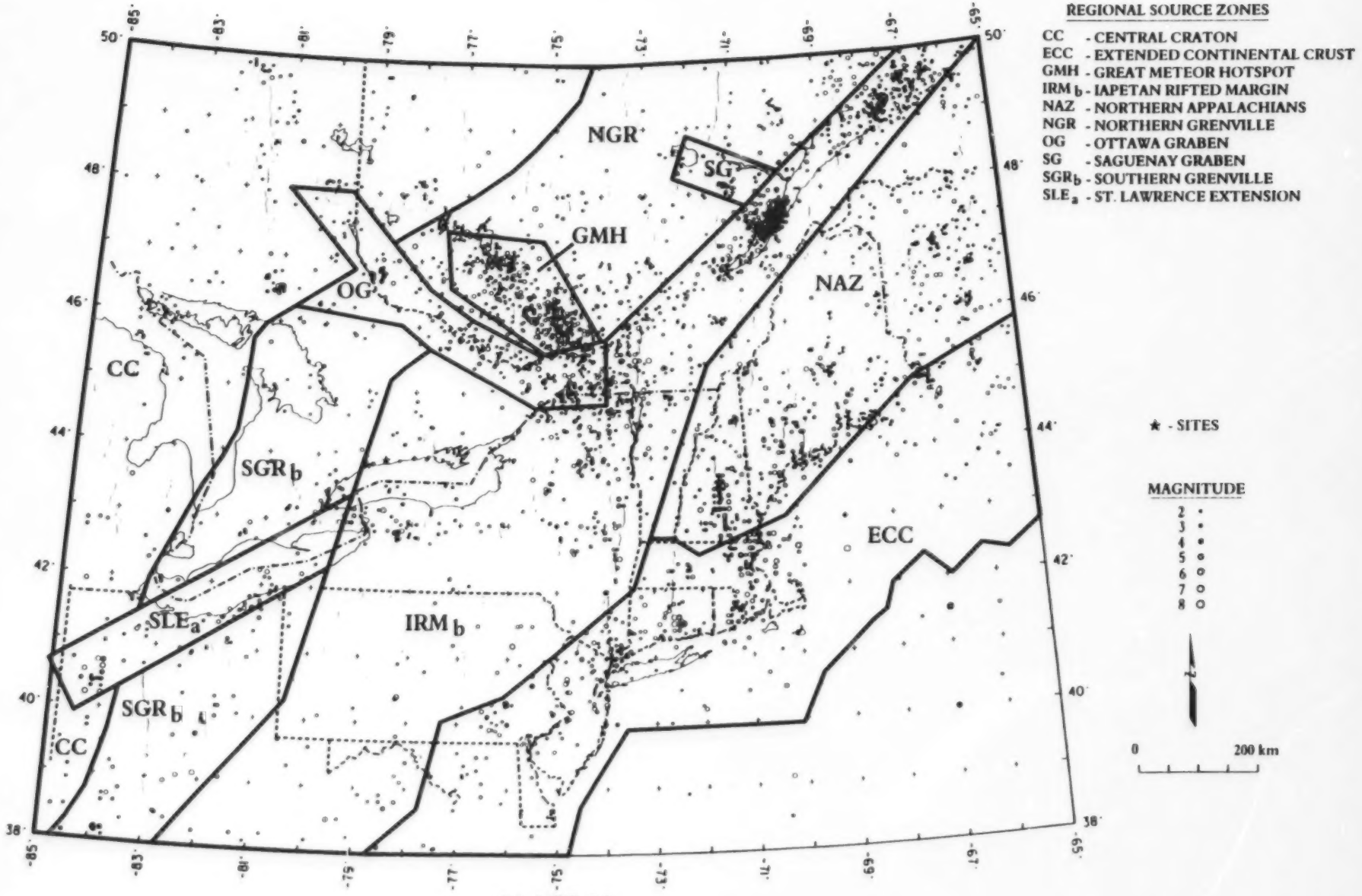

PLATE 5A

REGIONAL SOURCE ZONES FOR CMBBZ IRM/SGR BOUNDARY
WITH NARROW ST. LAWRENCE EXTENSION

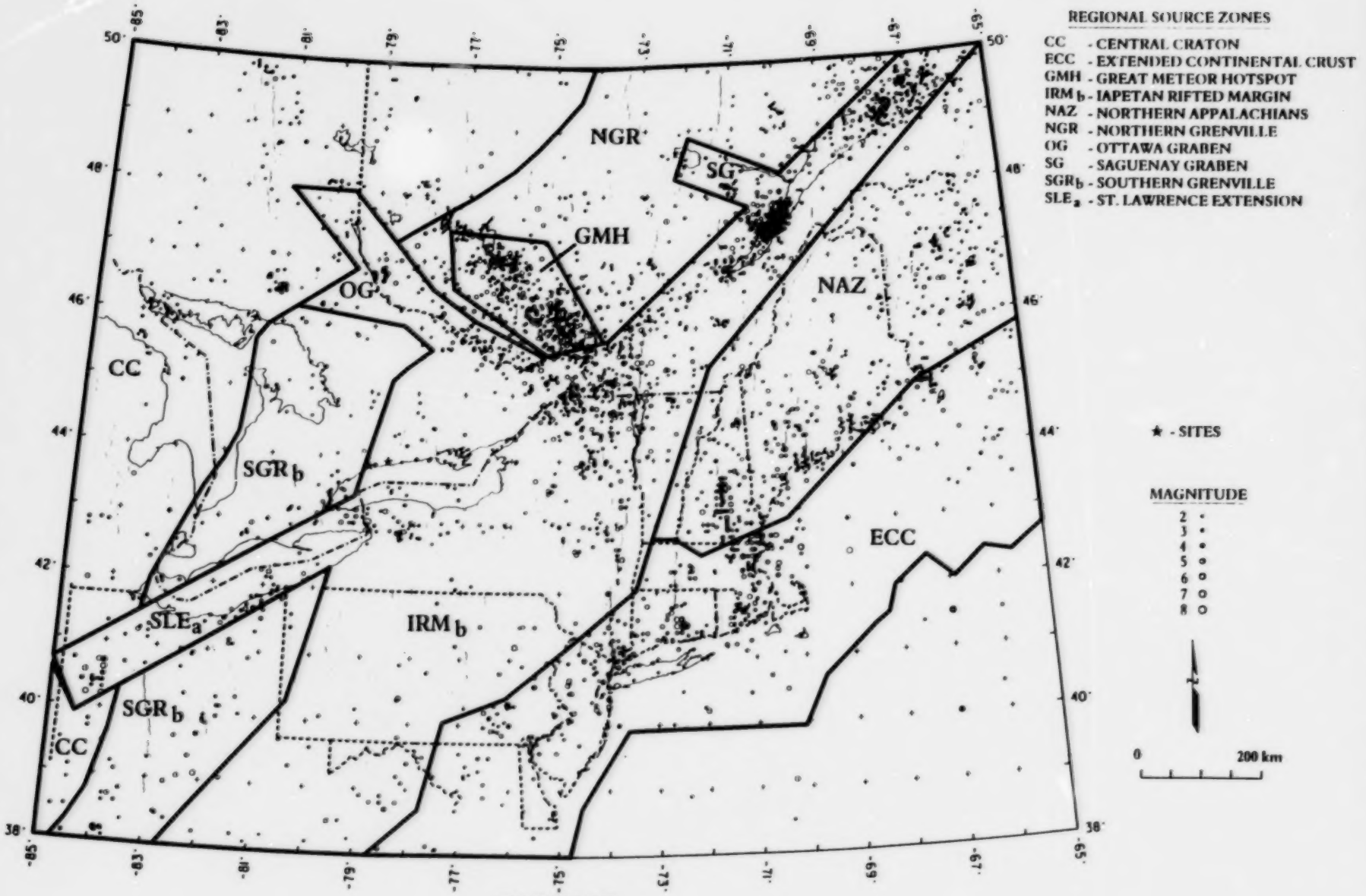

PLATE 5B

REGIONAL SOURCE ZONES FOR CMBBZ IRM/SGR BOUNDARY WITH NARROW ST. LAWRENCE EXTENSION

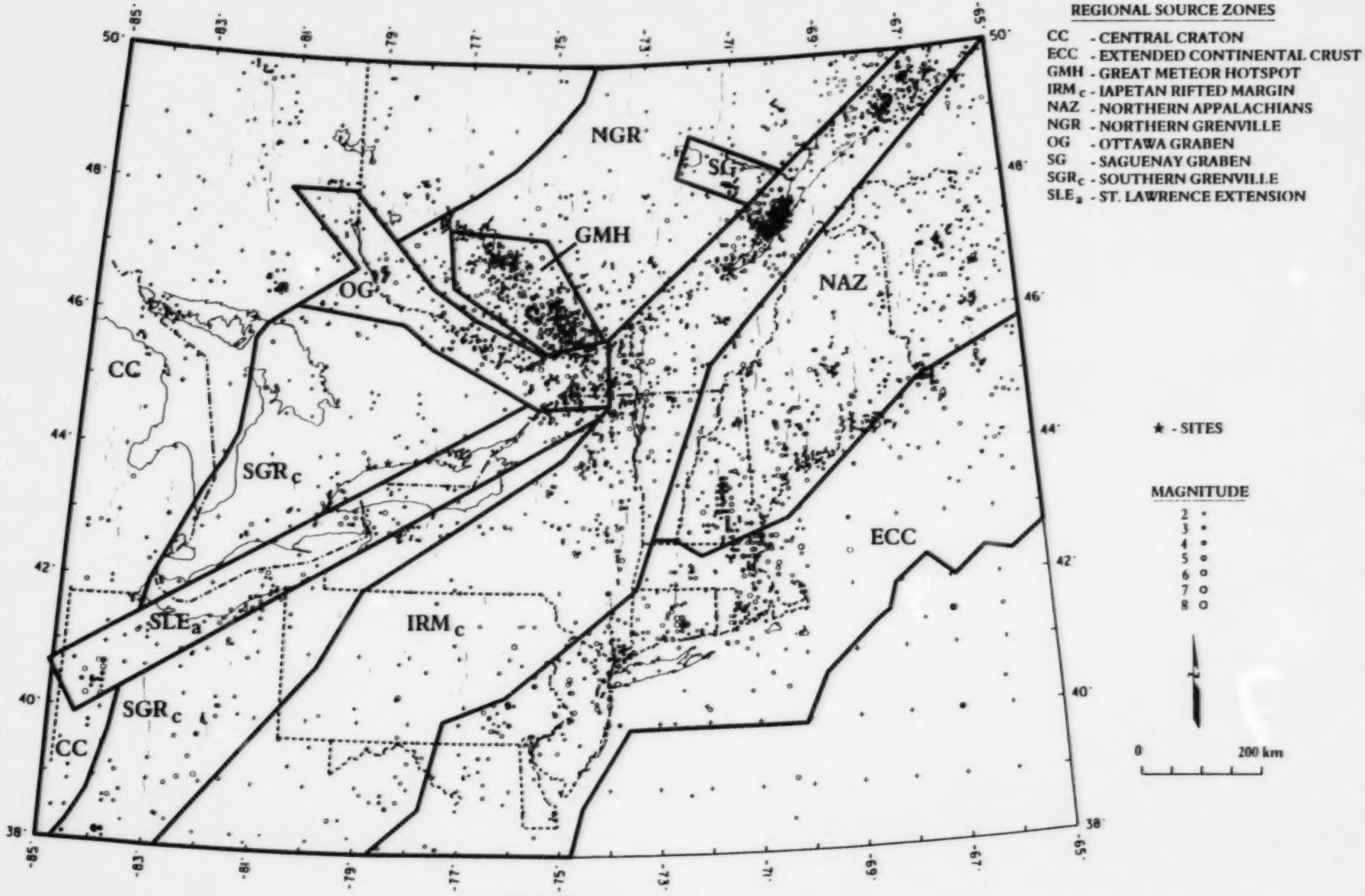

PLATE 6A

REGIONAL SOURCE ZONES FOR WESTERN ADIRONDACK UPLIFT IRM/SGR BOUNDARY WITH NARROW ST. LAWRENCE EXTENSION

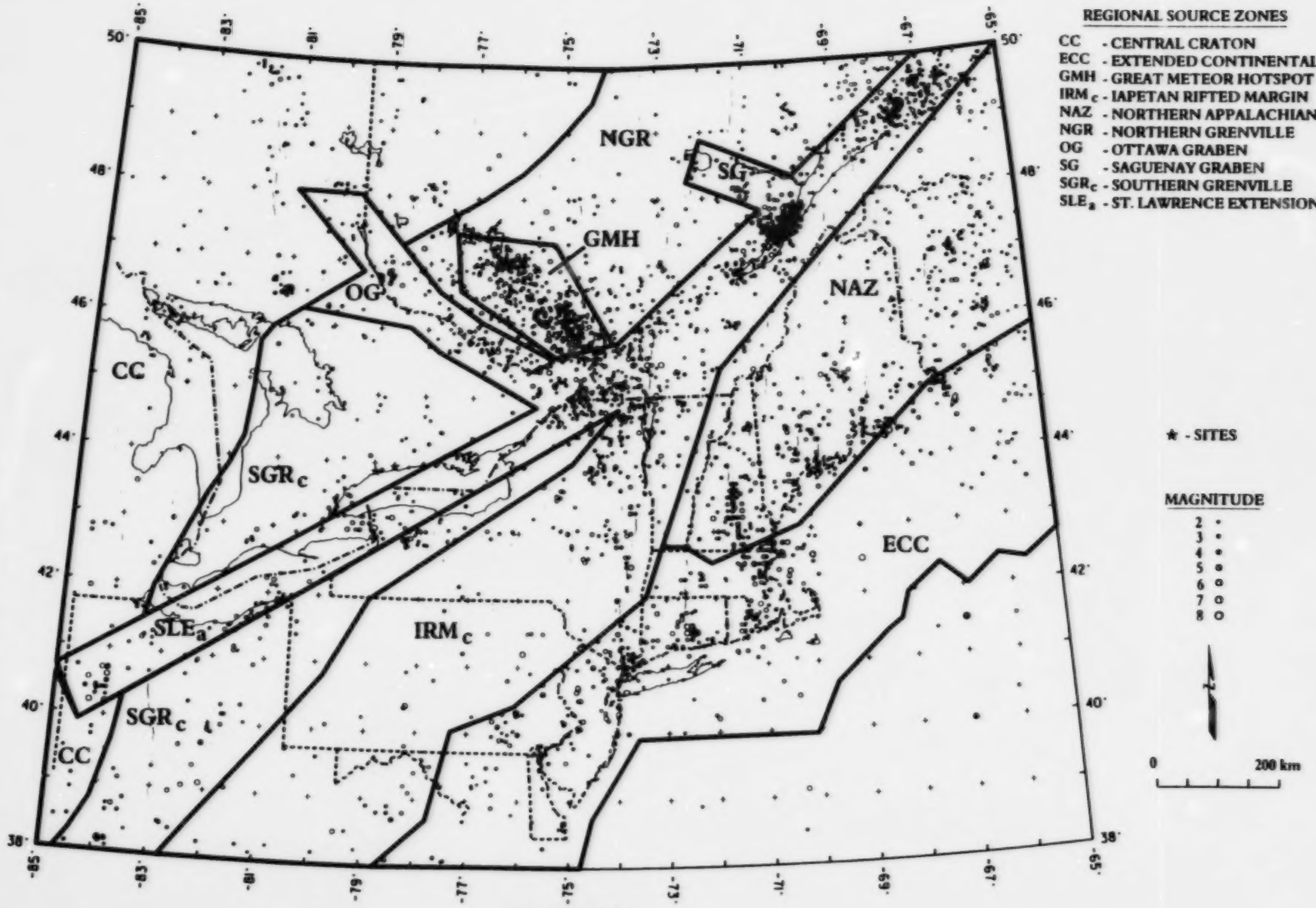

PLATE 6B

REGIONAL SOURCE ZONES FOR WESTERN ADIRONDACK UPLIFT IRM/SGR BOUNDARY WITH NARROW ST. LAWRENCE EXTENSION

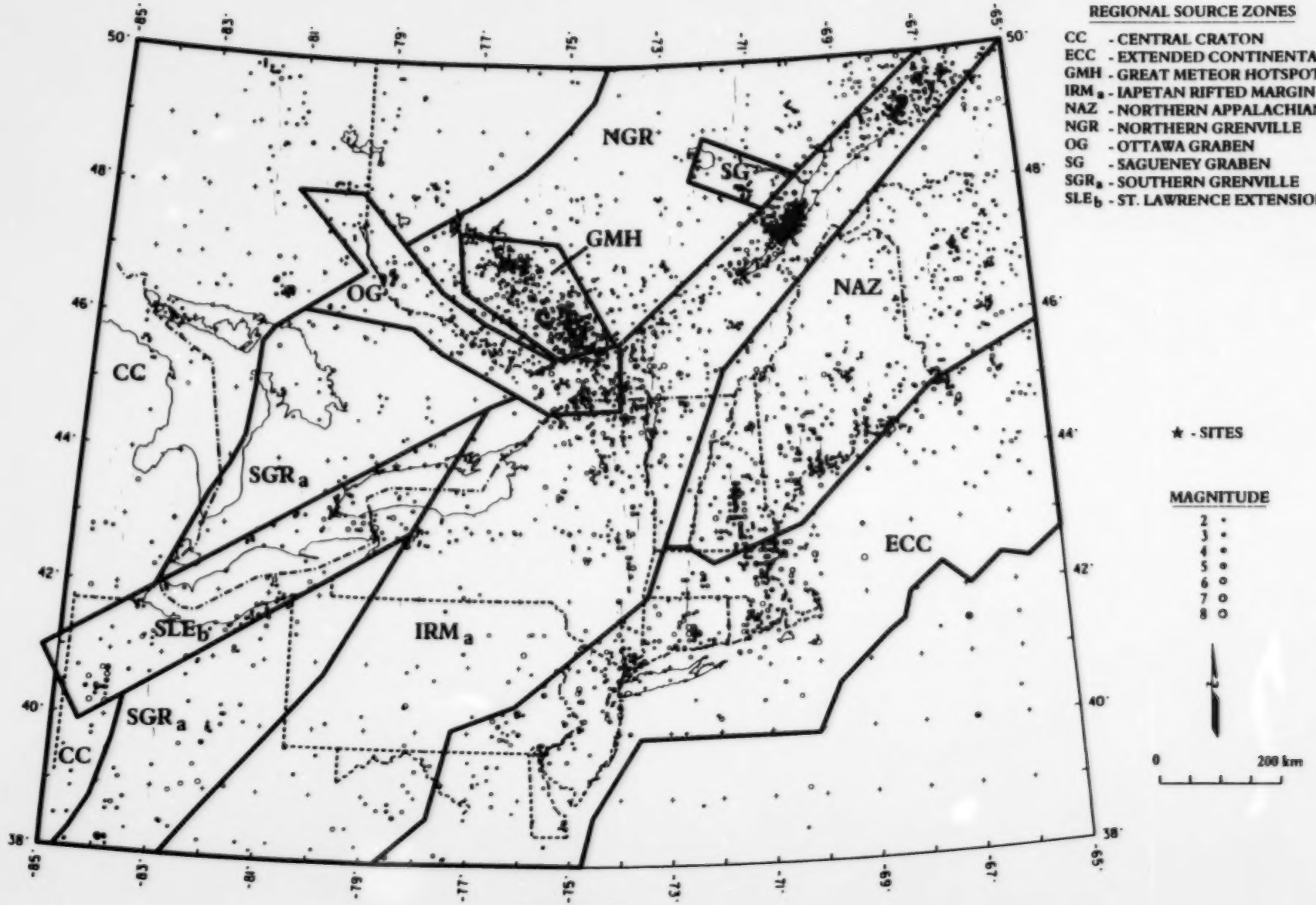

PLATE 7A

REGIONAL SOURCE ZONES FOR WHEELER IRM/SGR BOUNDARY WITH WIDE ST. LAWRENCE EXTENSION

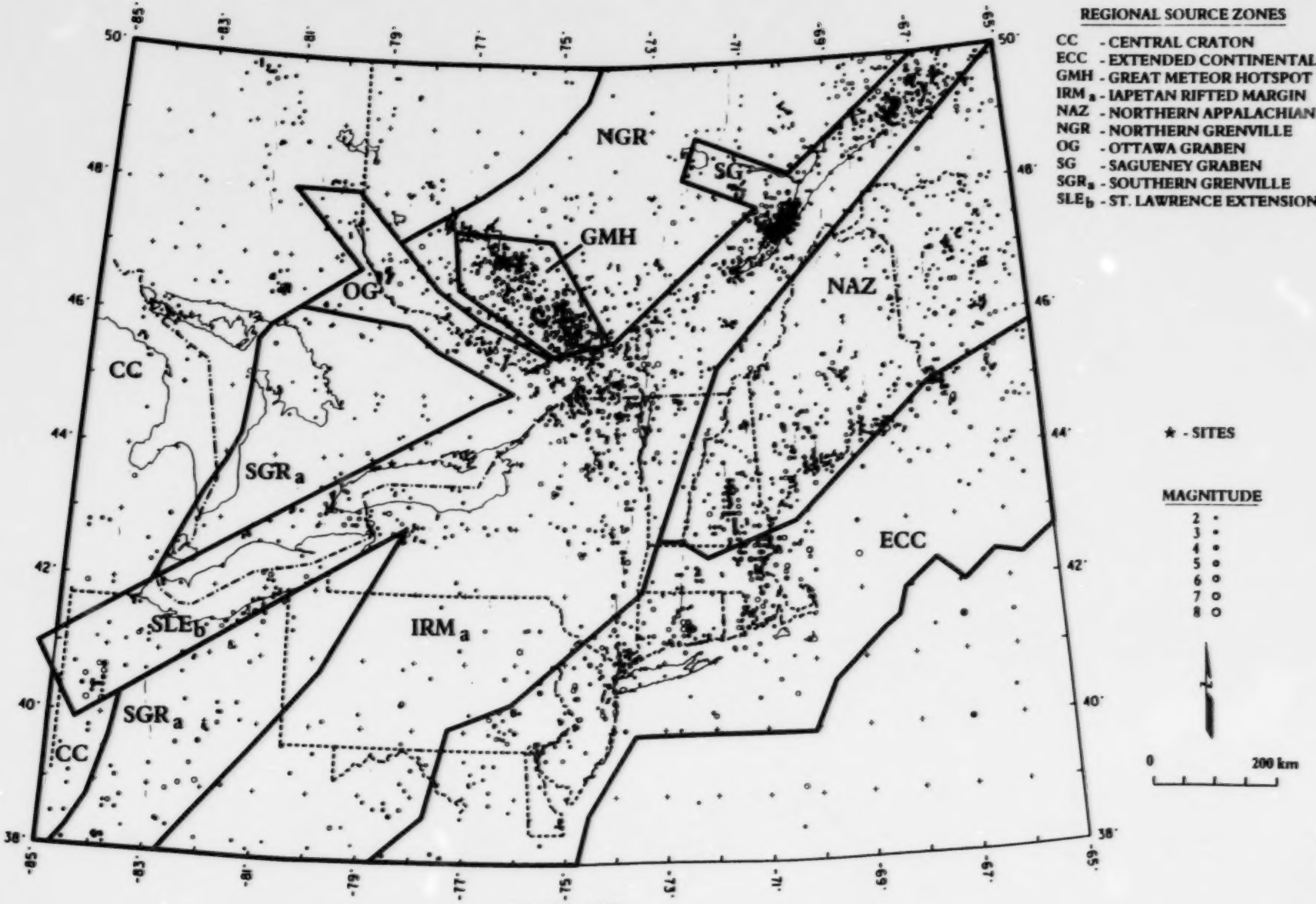

PLATE 7B

REGIONAL SOURCE ZONES FOR WHEELER IRM/SGR BOUNDARY WITH WIDE ST. LAWRENCE EXTENSION

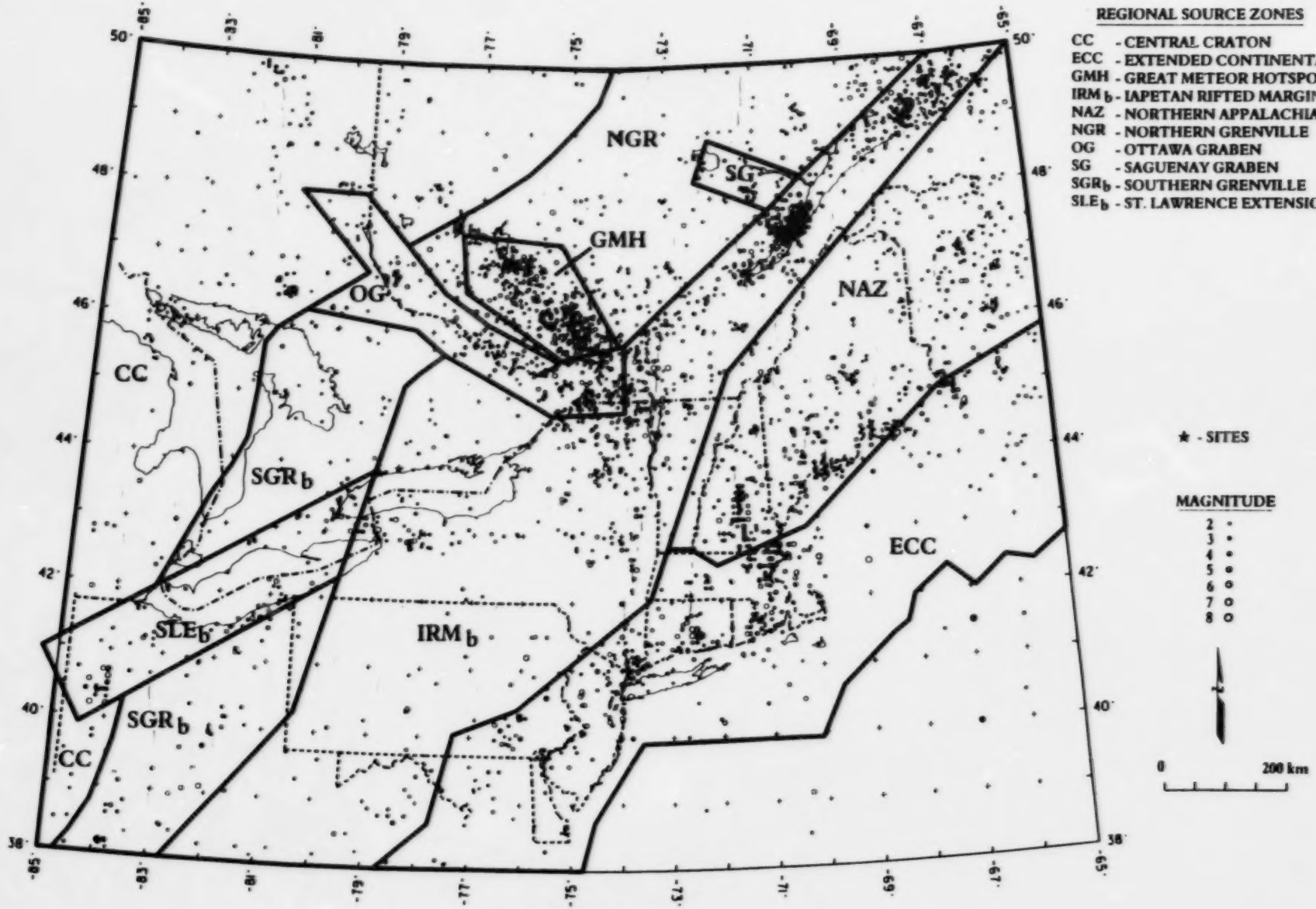

PLATE 8A

REGIONAL SOURCE ZONES FOR CMBBZ IRM/SGR BOUNDARY WITH WIDE ST. LAWRENCE EXTENSION

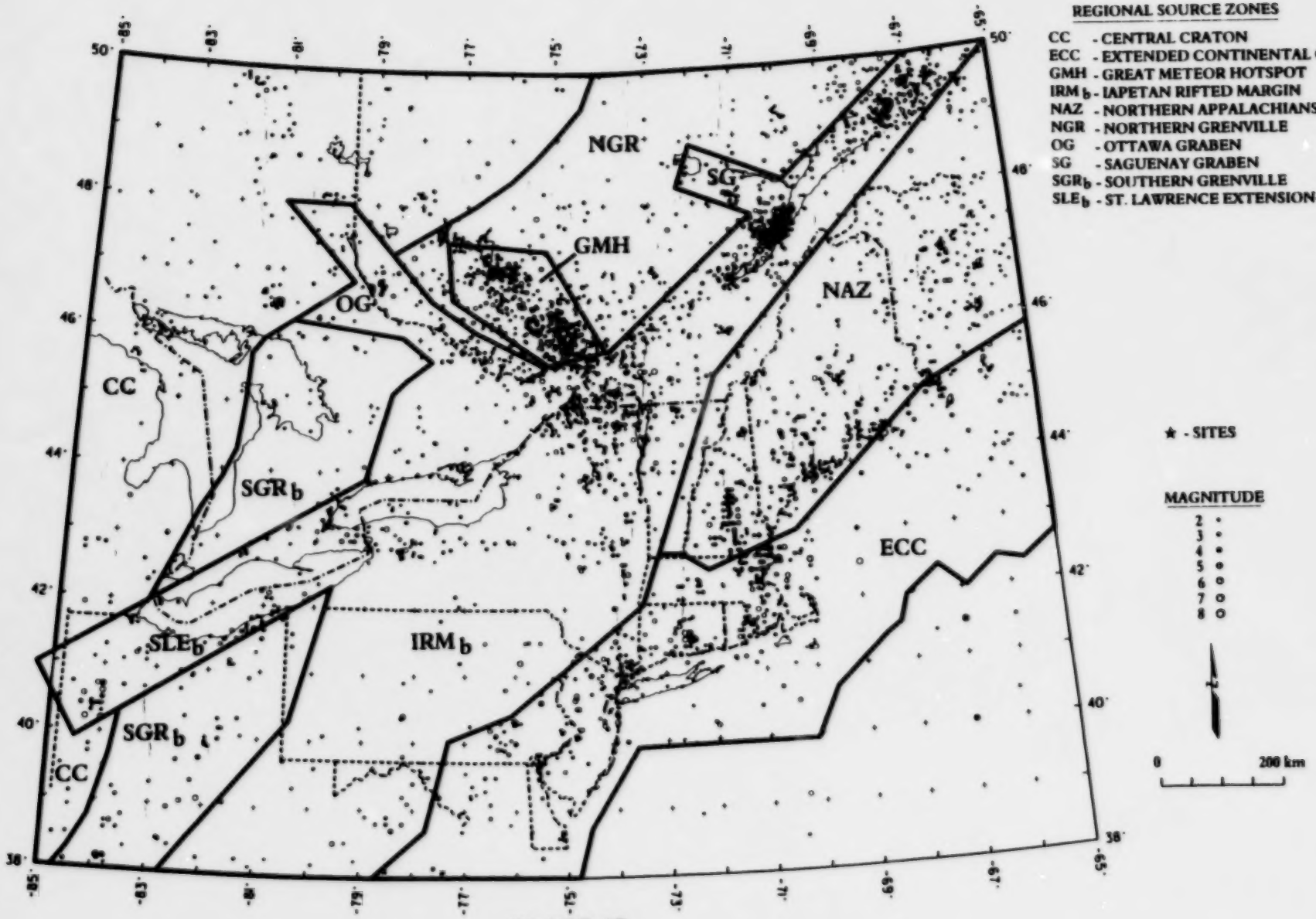

PLATE 8B

REGIONAL SOURCE ZONES FOR CMBBZ IRM/SGR BOUNDARY WITH WIDE ST. LAWRENCE EXTENSION

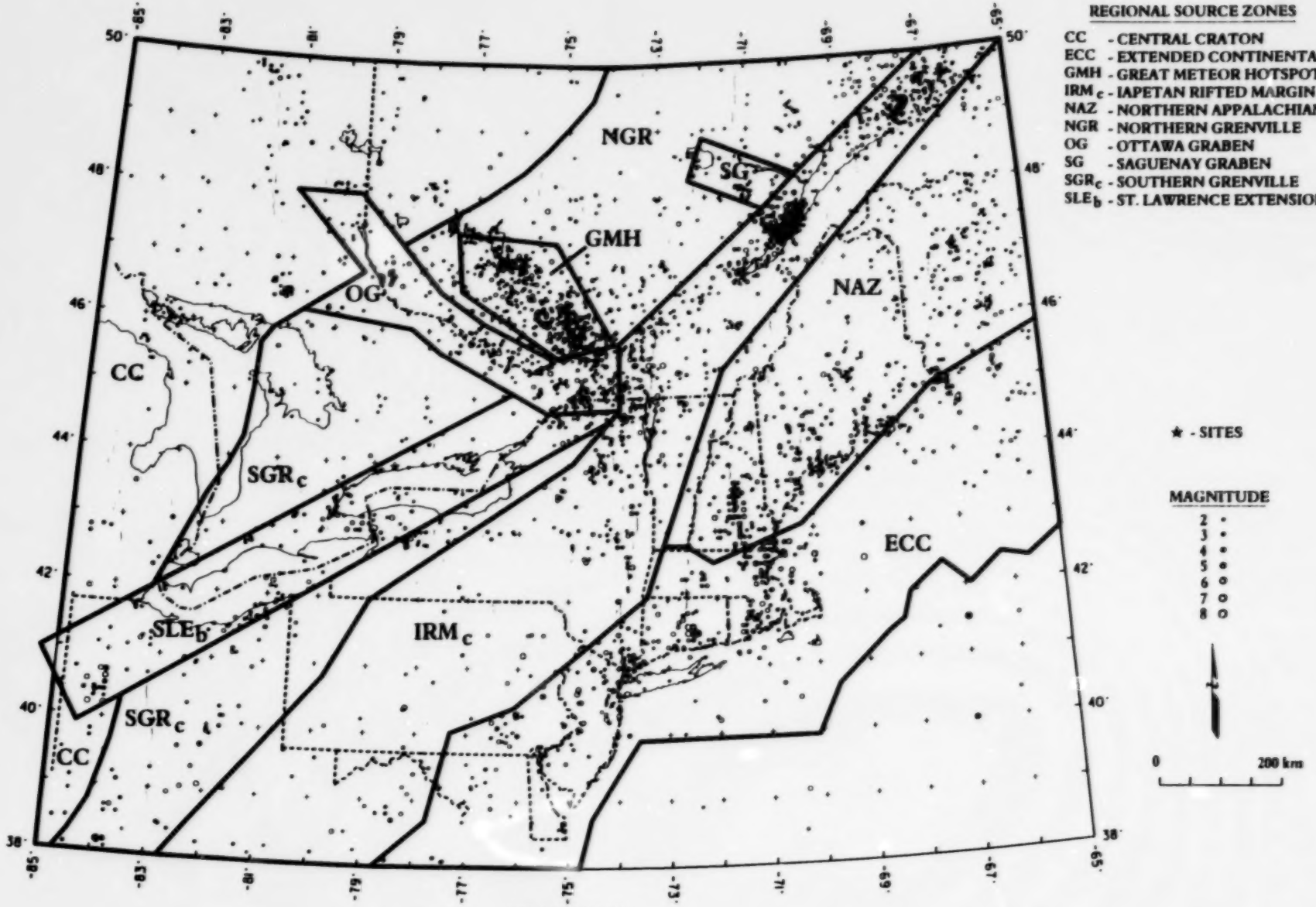

PLATE 9A

REGIONAL SOURCE ZONES FOR WESTERN ADIRONDACK UPLIFT IRM/SGR BOUNDARY WITH WIDE ST. LAWRENCE EXTENSION

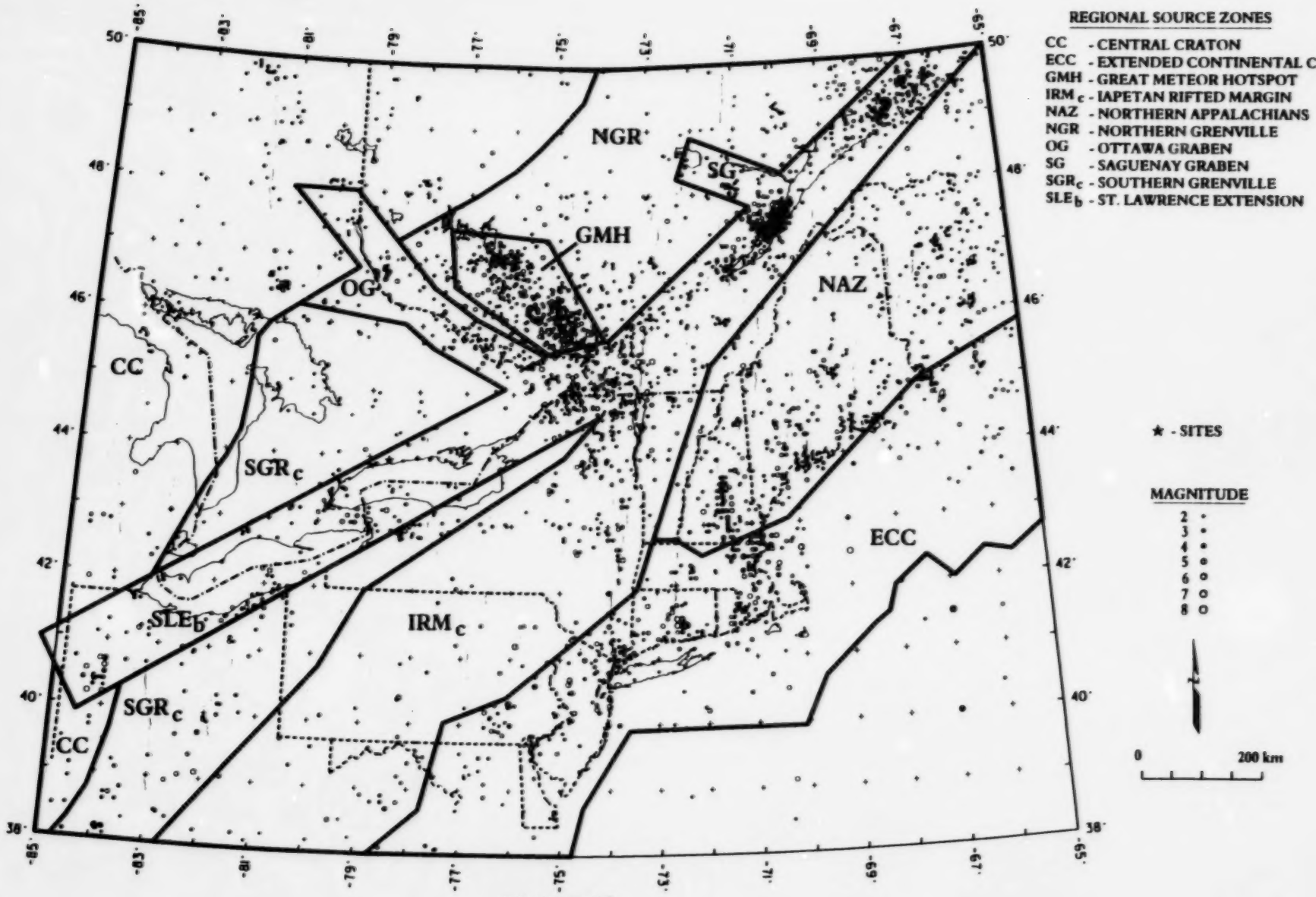

PLATE 9B

REGIONAL SOURCE ZONES FOR WESTERN ADIRONDACK UPLIFT IRM/SGR BOUNDARY WITH WIDE ST. LAWRENCE EXTENSION

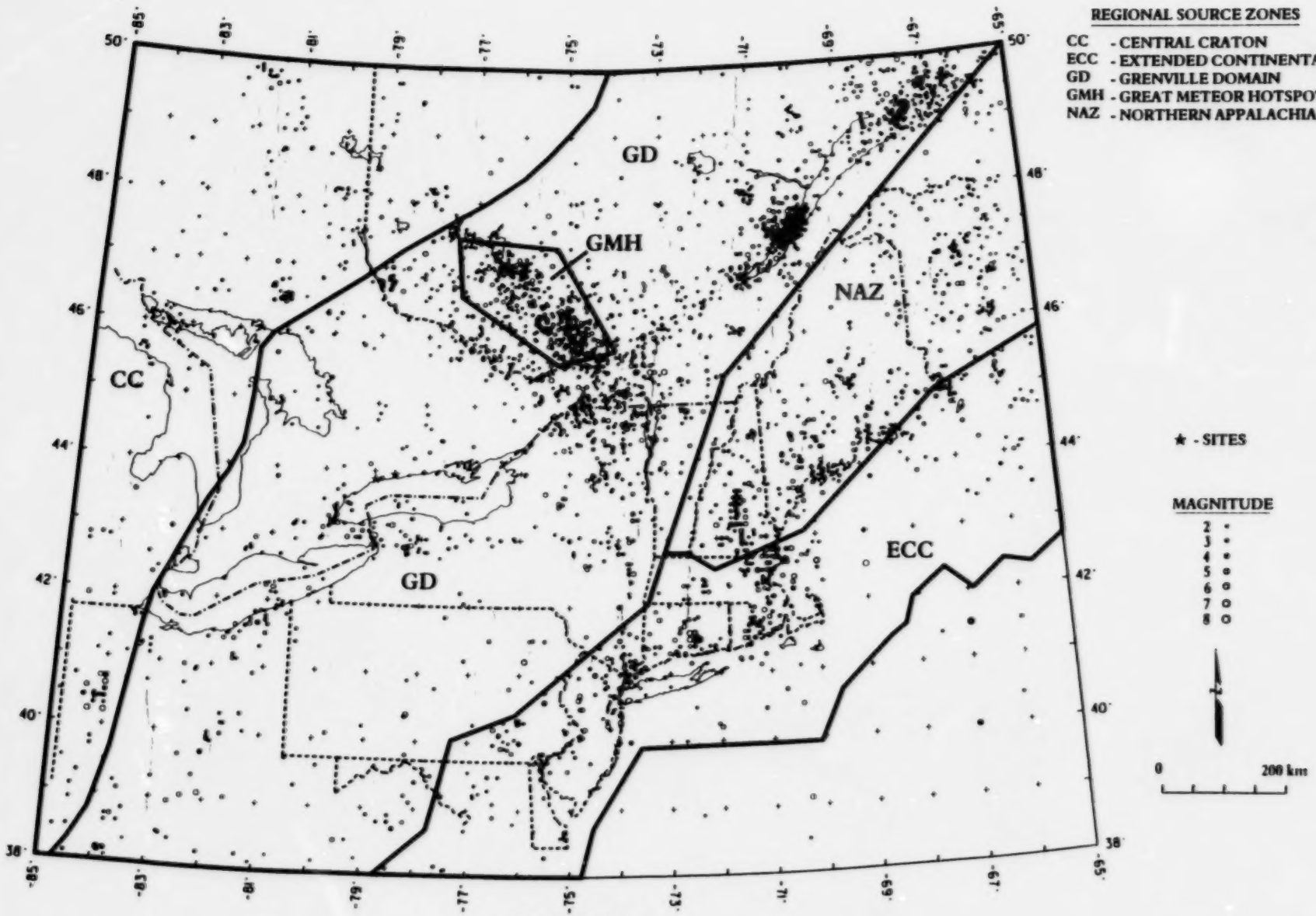

PLATE 10

REGIONAL SOURCE ZONES FOR NO IRM/SGR BOUNDARY

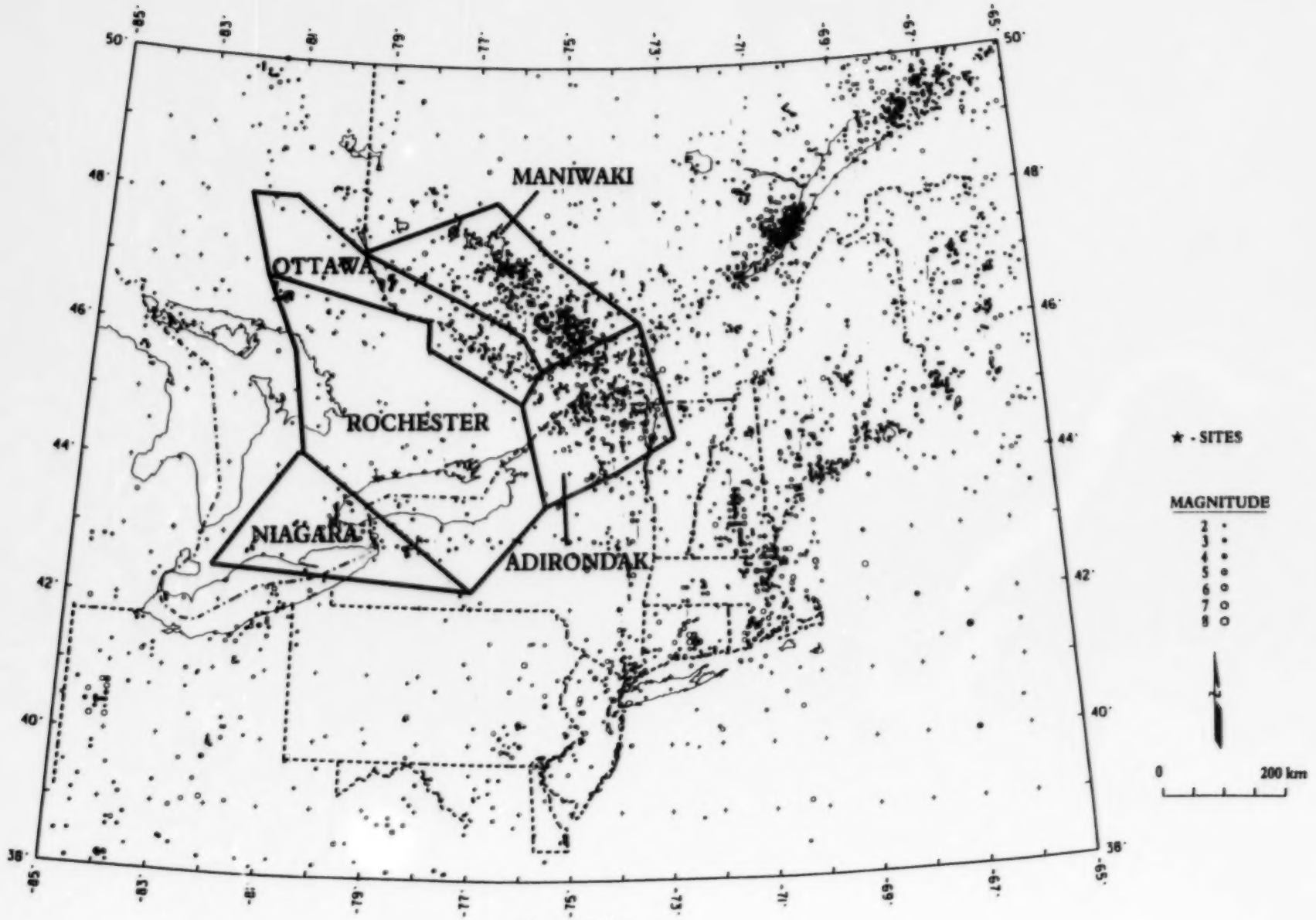

PLATE 11

REGIONAL COMPOSITE SOURCE ZONES (FROM ATKINSON, 1990)

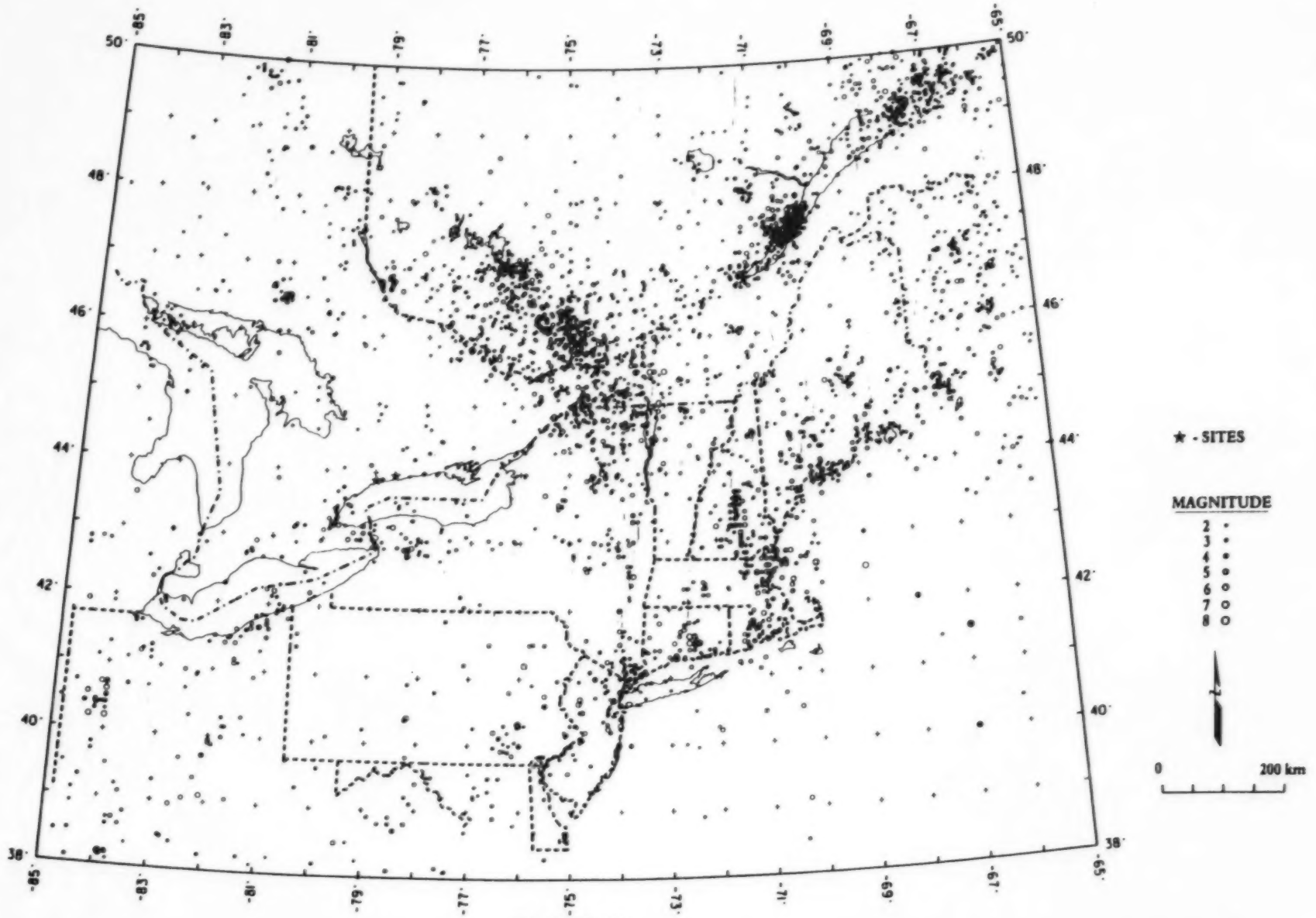

PLATE 12
REGIONAL SEISMICITY

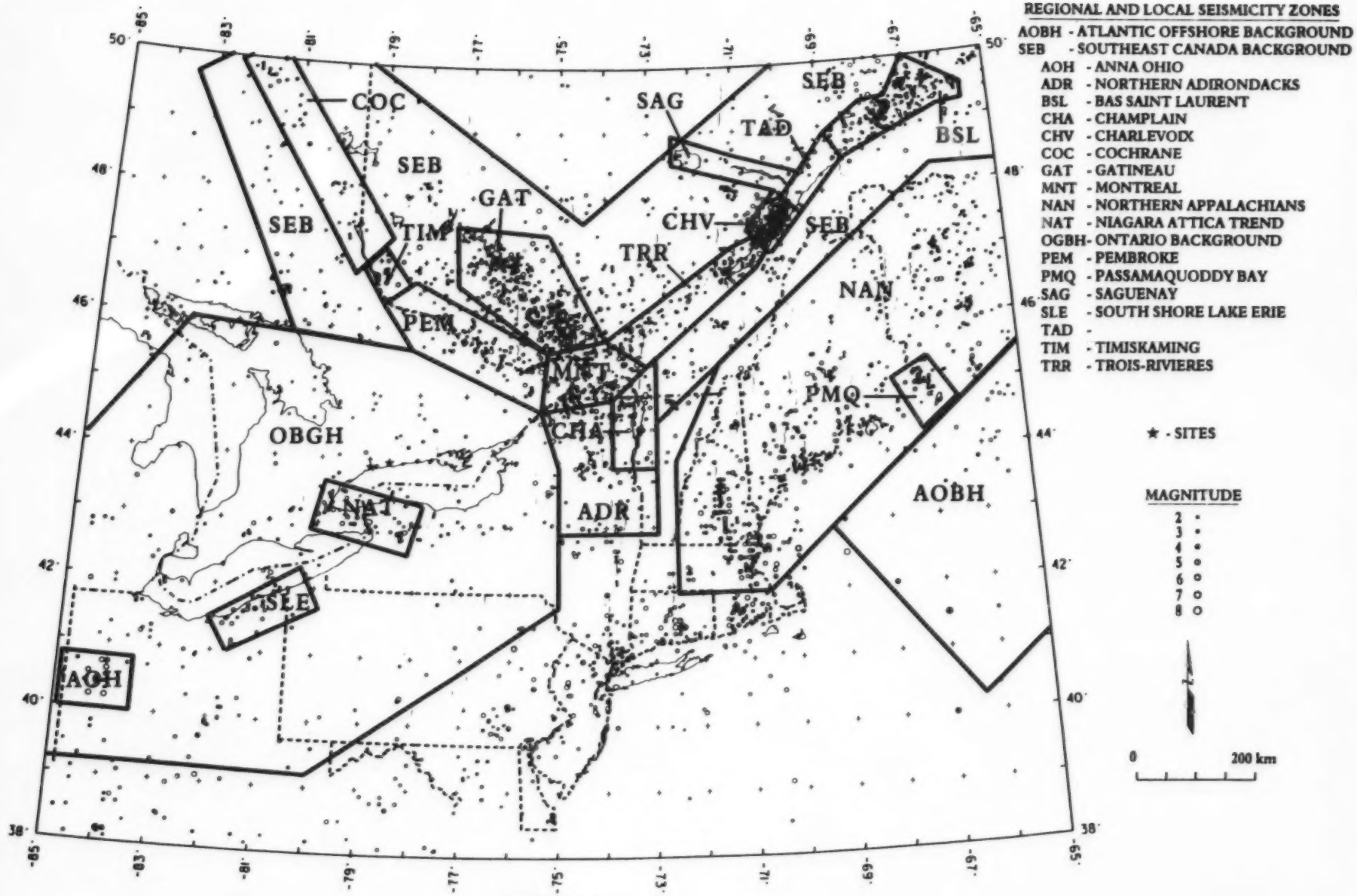

PLATE 13

REGIONAL AND LOCAL SEISMICITY ZONES (FROM ADAMS AND OTHERS, 1996)

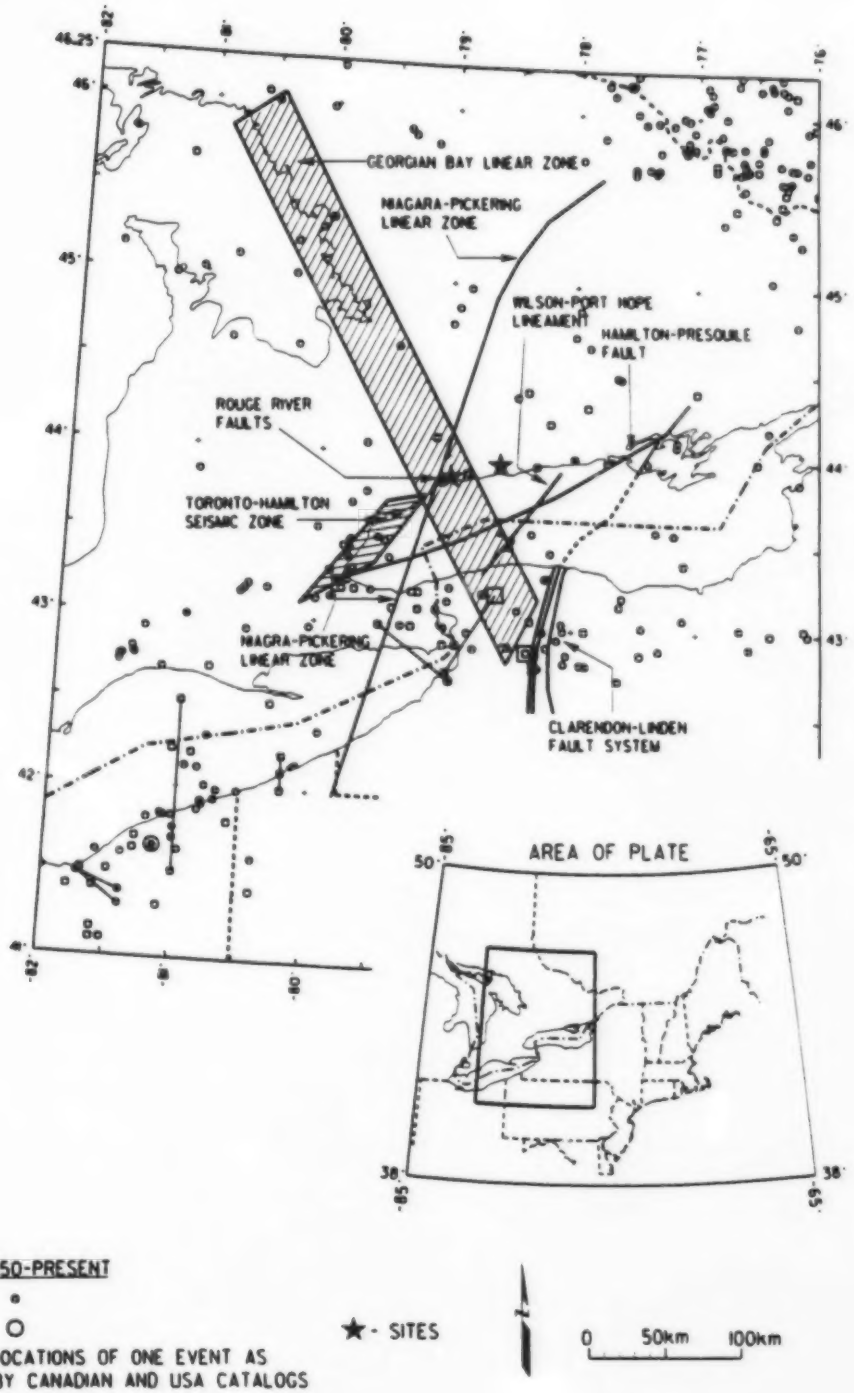

PLATE 14
LOCAL SOURCE MODEL 1

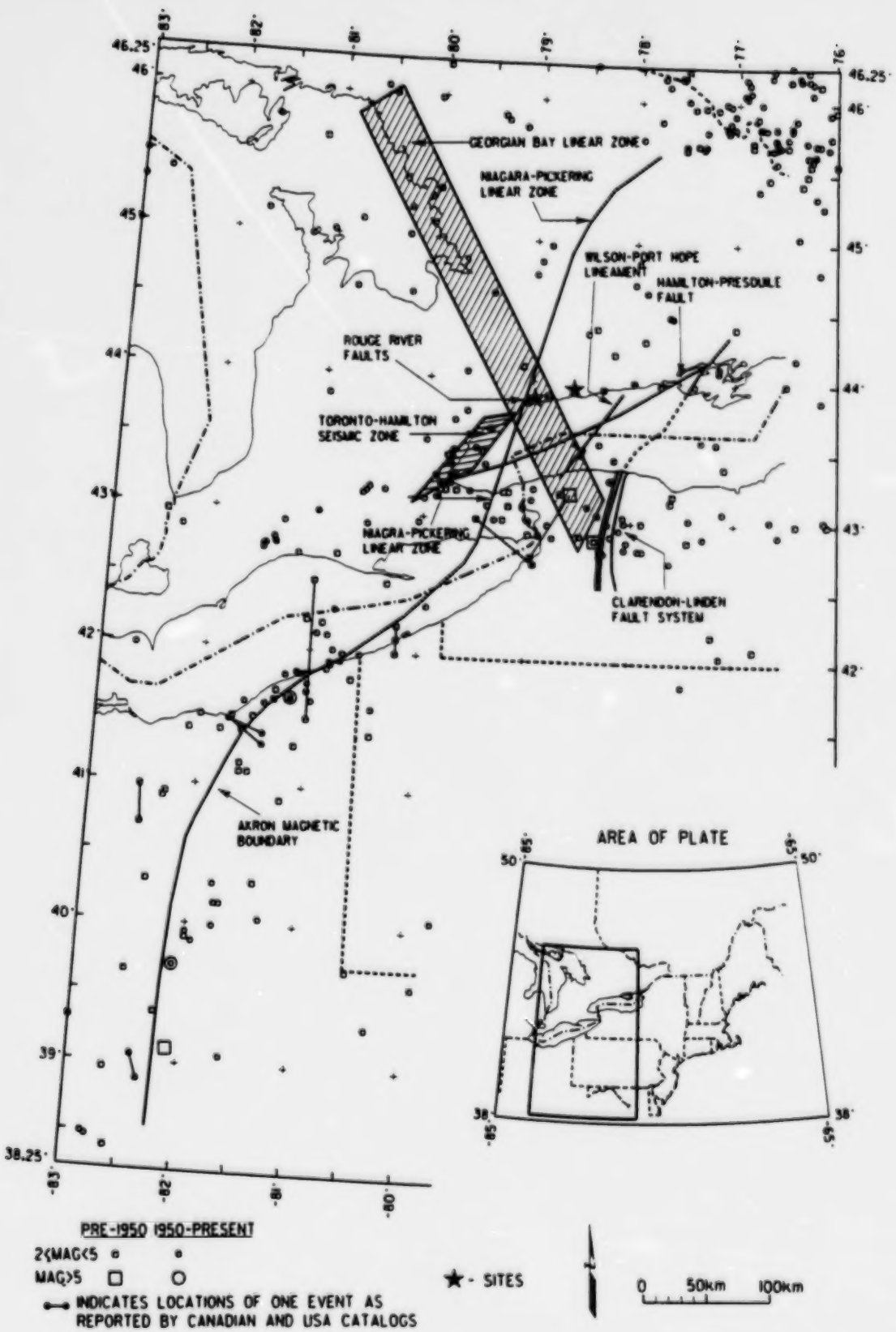

PLATE 15
LOCAL SOURCE MODEL 2